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Plio-Pleistocene geology of the Central Cagayan Valley, Northern Luzon, Philippines

by

Mark Evan Mathisen

A Dissertation Submitted to the
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For the Graduate College

Iowa State University
Ames, Iowa

1981

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INTRODUCTION

The Plio-Pleistocene terrestrial sediments of the central part of the Cagayan Valley, Northern Luzon, the Philippines, have been the focus of numerous archaeological and paleontological studies. Preliminary investigations by von Koenigswald (1958) and the Philippine National Museum (Fox, 1971; Lopez, 1972, Fox and Peralta, 1974) have documented the occurrence of Paleolithic pebble-cobble tools of the Asian chopper-chopping tool tradition, flakes, and fossilized remains of a Pleistocene vertebrate fauna. The artifacts and fossils have been found at sixty-eight sites on the eroded tops and slopes of grassy, often gravel covered anticlinal hills. Most of the artifacts are surface finds of unknown age that have not, as yet, been directly associated with the vertebrate fossils (Fox and Peralta, 1974) which are middle Pleistocene in age.

In China (Black, 1933; Movius, 1949) and Indonesia (DeTerra, 1943; von Koenigswald and Ghosh, 1973) the same chopper-chopping tool tradition has been dated by direct association with an extinct middle Pleistocene fauna and *Homo erectus*. Previous workers in the Cagayan Valley (Fox and Peralta, 1974) have postulated that there may also be a direct association between the Philippine artifacts and extinct middle Pleistocene fauna indicating that *Homo erectus* inhabited the Philippines during the middle Pleistocene.

To better understand the significance of the Paleolithic artifacts and Pleistocene fauna with regard to Philippine and Southeast Asian prehistory, a cooperative project between archaeologists and geologists was organized and initiated in 1978. Field studies utilizing an interdisciplinary approach are being conducted to (1) define the Plio-Pleistocene terrestrial sequence in the Cagayan Valley basin, (2) determine the relationship between the artifacts and Pleistocene fauna, (3) interpret the age of the artifacts, (4) identify specific animal species, and (5) interpret the Plio-Pleistocene environments of the valley (Shutler and Mathisen, 1979).

This report describes the Plio-Pleistocene geology of the central part of the Cagayan Valley. This is an area of over 2,000 square km, between 121°25' and 121°50' east longitude and 17°20' and 17°45' north latitude, which encompasses the towns of Peñablanca, Tuguegarao, Solana, Rizal, and Tabuk (Figure 1). The principal objectives of the report are:

1. to provide a stratigraphic framework for the accurate documentation and age determination of artifacts and fossils collected;
2. to interpret the regional depositional history of the Plio-Pleistocene terrestrial sediments;
3. to interpret local paleoenvironments represented by sediments exposed at archaeological and paleontological sites; and
4. to reconstruct the Plio-Pleistocene history of the central Cagayan Valley.

Method of Study

Field studies were conducted for a total of seven months during the dry seasons of 1978, 1979, and 1980. Excursions were first made to various parts of the valley and the adjacent mountains to observe the regional geology. Thirty-five stratigraphic sections were then measured throughout the central Cagayan Valley to provide a basis for interpreting the Plio-Pleistocene geology, stratigraphy, and paleoenvironments. Fourteen of the

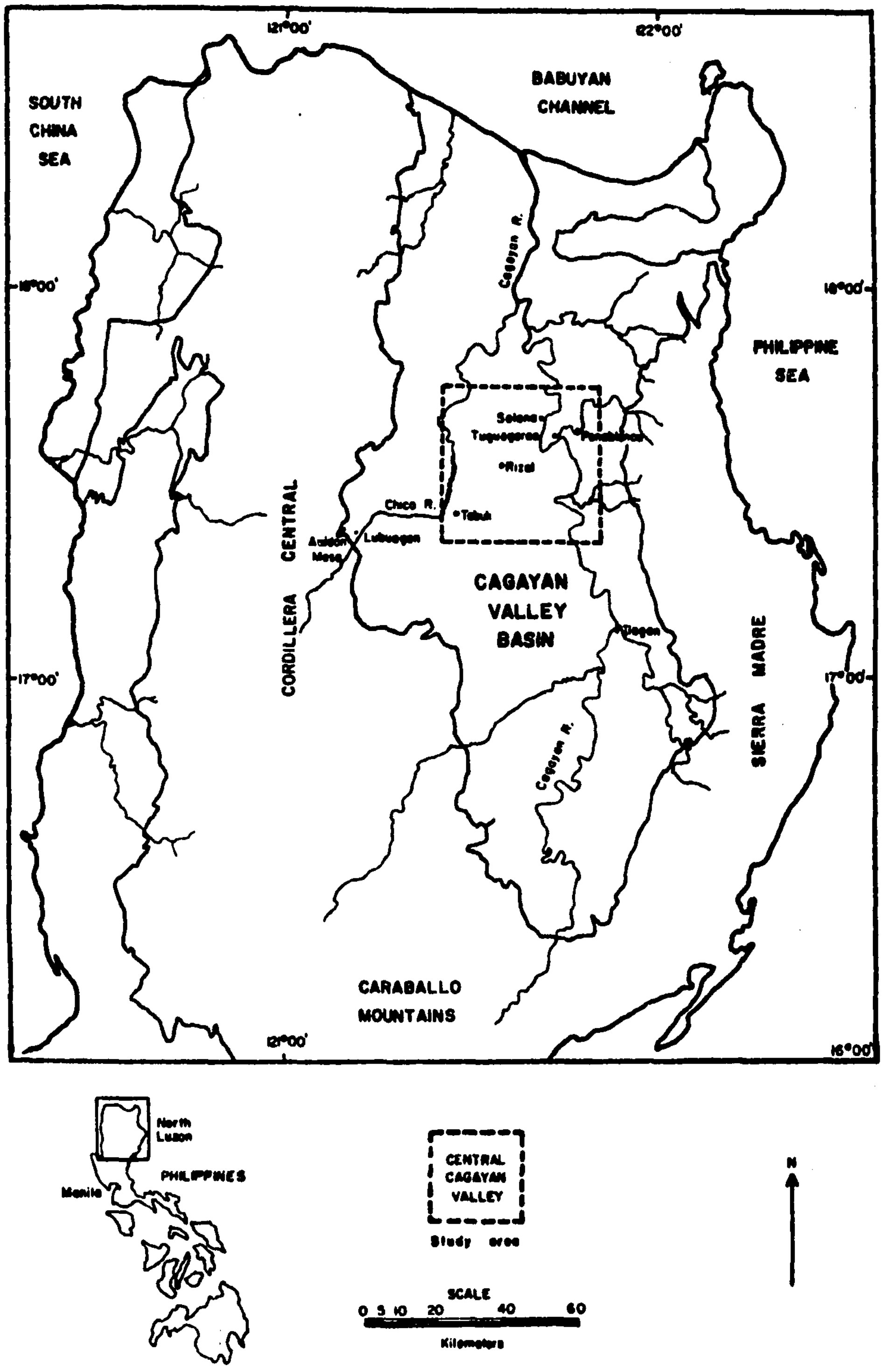


Figure 1. Index map of the central Cagayan Valley

sections were measured in the Cabalwan Anticline area where most of the Philippine National Museum archaeologic and paleontologic sites occur. Fundamental rock stratigraphic units (formations) were defined, and laterally continuous pyroclastic deposits were delineated as marker beds. A geologic map of the central Cagayan Valley was prepared utilizing 1:50,000 topographic maps as base maps. Some data are inferred from previous maps by Durkee and Pederson (1961) and Christian (1964) since the use of aerial photographs was restricted by the government and parts of the study area were inaccessible.

A variety of qualitative and quantitative laboratory procedures were followed to supplement the fieldwork. The composition of the sediments was studied using thin section, grain mount, X-ray diffraction, and scanning electron microscope methods. Textural parameters of the sediments that had not been significantly altered by diagenetic processes were calculated for samples analyzed by mechanical and sedimentation techniques. The laboratory methods are described in more detail in the petrology section of this report.

Climate and Vegetation

Seasons are not very pronounced in the Cagayan Valley, but the region is generally dry from January to April and wet during the remainder of the year. At Tuguegarao, the average annual rainfall for the period 1951-1970 was 1700 mm. The monthly average rainfall reaches a peak in November with 280 mm (Wasson and Cochrane, 1979). Rainfall is generally higher in the adjacent mountains where a large portion of the Cagayan River catchment lies. Northern Luzon is subject to northeast tradewinds from January to

April, east winds from April to July, southeast tradewinds from July to October, and east winds from October to December (Durkee and Pederson, 1961). During the latter months of the year, the wind system is complicated by typhoons which generally pass from east to west across Luzon (Dickerson, 1928) and often affect the Cagayan Valley. Temperatures average 27.7°C at Tuguegarao with a range of average monthly temperatures from 24.5°C in January to 30.5°C in May (Wasson and Cochrane, 1979).

The vegetation of the central Cagayan Valley has been significantly modified due to agricultural practices. Slash and burn agriculture has reduced much of the primary forest (probably Dipterocarp; Dickerson, 1928) to isolated patches along streams and gullies. Grasslands now predominate in the upland areas while the lowland areas are used for rice production.

Previous Work

Most previous geologic investigations of the Cagayan Valley were conducted to evaluate the hydrocarbon potential of the sediments. Corby et al. (1951), Irving (1952), Kleinpell (1954), Vergara et al. (1959), and Durkee and Pederson (1961) completed the first reconnaissance studies of the Cagayan Valley sediments and established the stratigraphic nomenclature. The tectonic history of the Cagayan Valley basin was then interpreted by Christian (1964). More recent geologic investigations of the sediments have been conducted by the Philippine National Oil Company (Tamesis, 1976; Caagusan, 1978, 1980). All of these studies have been concentrated on the Miocene marine sediments which have the greatest oil potential and have only briefly described the Plio-Pleistocene sediments.

More detailed geologic investigations of the Plio-Pleistocene sediments were initiated by the Philippine National Museum in the early 1970s. Lopez (1971, 1972) made a preliminary interpretation of the geology and paleontology of the Pleistocene deposits. Difficulties in age documentation and correlation of strata led to a moratorium on all archaeologic field work until the Pleistocene geology of the area was better understood (Shutler and Mathisen, 1979). This investigation was initiated in 1978 to provide the geologic background necessary for interpreting archaeologic and paleontologic sites (Mathisen and Vondra, 1978; Vondra et al., 1981). Additional investigations have since been initiated which describe the geomorphology (Wasson and Cochrane, 1979), Mio-Pliocene transitional marine sediments (Kvale, 1981), and various characteristics of the Pleistocene pyroclastic deposits (Burggraf, 1981; Ross, 1981).

Geologic Setting

The Cagayan Valley is a major north-south trending intermontane basin approximately 250 km long and 80 km wide which developed in the mobile belt bordering the Asian mainland. The tectonic elements of this belt in the Luzon-Taiwan region are shown in Figure 2. The Manila trench occurs to the west of Luzon while the Quezon trench occurs to the east. A faulted submarine ridge, the North Luzon ridge (Mammerickx et al., 1976; Karig, 1973) extends between Luzon and Taiwan to the north. The Cagayan basin itself is bordered to the north by the sea and by mountain ranges to the east, south, and west (Figure 1). The Paleogene Sierra Madre range occurs to the east and is a volcanic arc composed of sialic basement (DeBoer et al., 1980), intermediate (andesitic) igneous rocks, diorite intrusives,

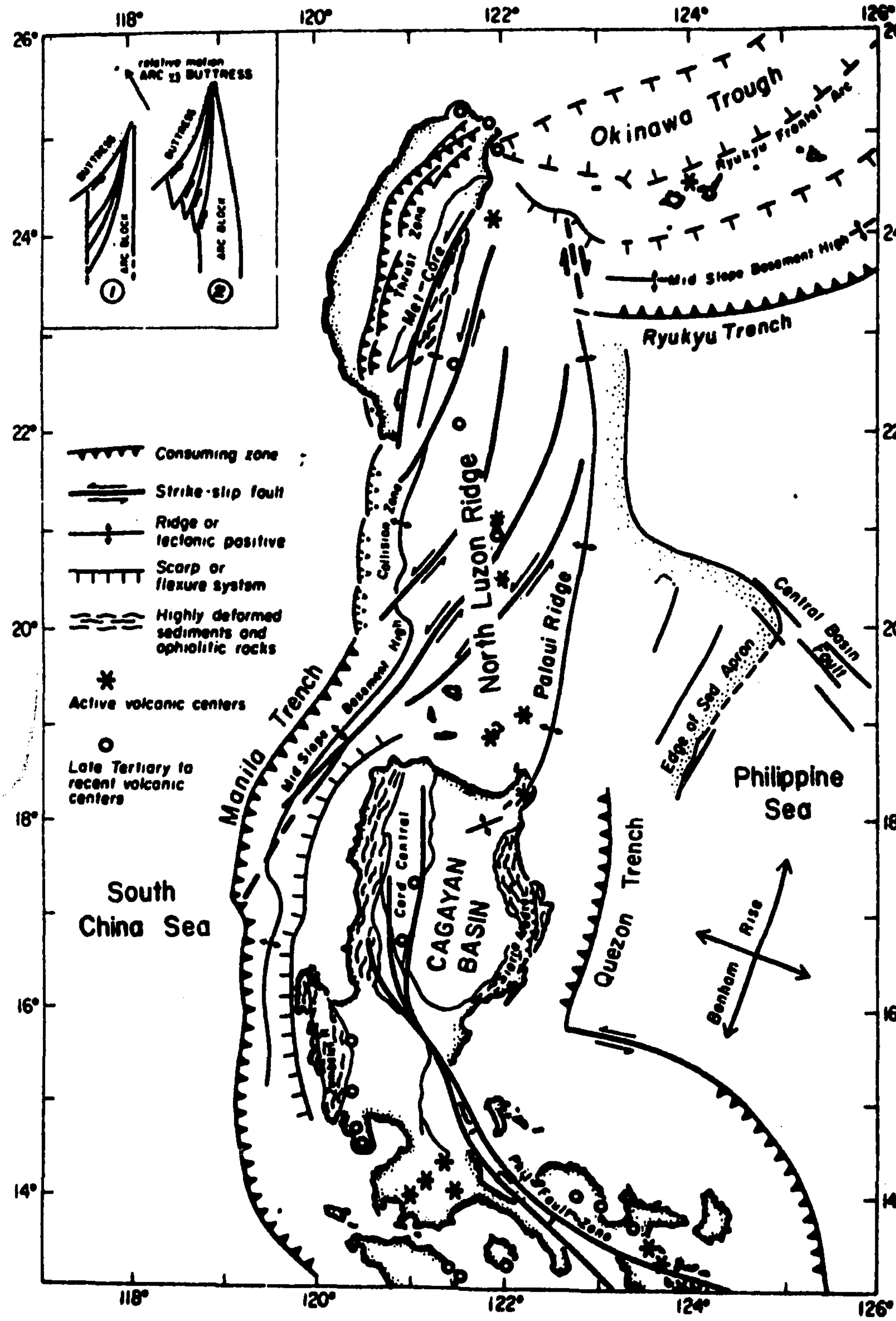


Figure 2. Tectonic map of the Luzon-Taiwan region. (The inset schematically shows a possible mechanical solution to explain the internal deformation within the arc system (after Karig, 1973))

metavolcanics, metasediments (Durkee and Pederson, 1961), and ophiolites that consist of spilite and associated chert (Caagusan, 1980). To the south, the late Eocene Caraballo range is formed by schists and folded volcanic and pyroclastic rocks metamorphosed to the pumpellyite-prehnite and green schist facies (Hashimoto et al., 1980). The Neogene Cordillera Central borders the Cagayan basin to the west and is a volcanic arc composed of mafic to intermediate plutonic rocks, basalts, andesites, metasediments, and silicic intrusives and extrusives (Durkee and Pederson, 1961; Balce et al., 1980) of calc-alkaline to shoshonitic (?) composition (DeBoer et al., 1980).

Northern Luzon began to form as an island arc system in the Mesozoic along the margin of mainland China (Audley-Charles, 1978) and migrated to its present location in the Tertiary (DeBoer et al., 1980) (Figure 3). The Paleogene Sierra Madre represents the remains of an east facing arc which formed in response to westward subduction of Philippine basin crust below the Asian continental block in the early Tertiary (Figure 3a). In the late Oligocene, east-west spreading was initiated in the Parece-Vela basin (Karig, 1975) to the east of the Sierra Madre volcanic arc. This spreading may have interrupted the eastward motion of the Philippine plate (DeBoer et al., 1980) resulting in a reversal of the Sierra Madre arc polarity (Karig, 1973; Murphy, 1973; Bowin et al., 1978) and eastward subduction of South China sea floor beneath the Philippine basin plate and Sierra Madre arc (Figure 3b). The Cordillera Central volcanic arc then began to form in the late Oligocene and early Miocene as a result of subduction of the South China basin seafloor. The Cagayan basin also formed at this time (Figure 3c) as an interarc basin behind the active Cordillera Central. In the late

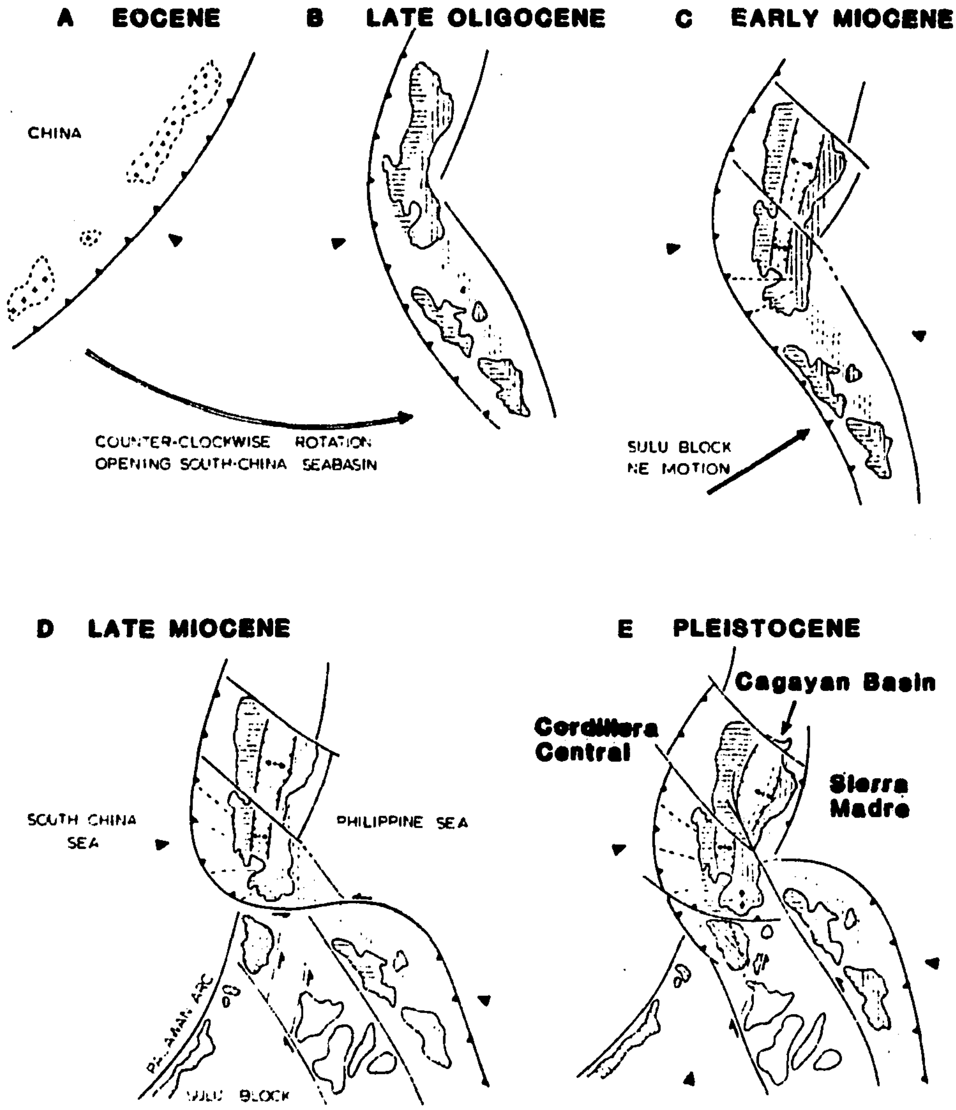


Figure 3. Island arc evolution of Northern Luzon (after DeBoer et al., 1980)

Miocene, the northeast motion of the Sulu block caused formation of a major shear fault in the Verde passage and formation of the Philippine trench (Figure 3d) (DeBoer et al., 1980). In the Pleistocene, westward motion of the Philippine basin plate resulted in oblique subduction along the Philippine trench, sinistral strike slip movement along the Philippine fault and reactivation of subduction along the Quezon trench (Figure 3e)

(DeBoer et al., 1980). The various events in the island arc evolution of Northern Luzon just described are summarized in Figure 4.

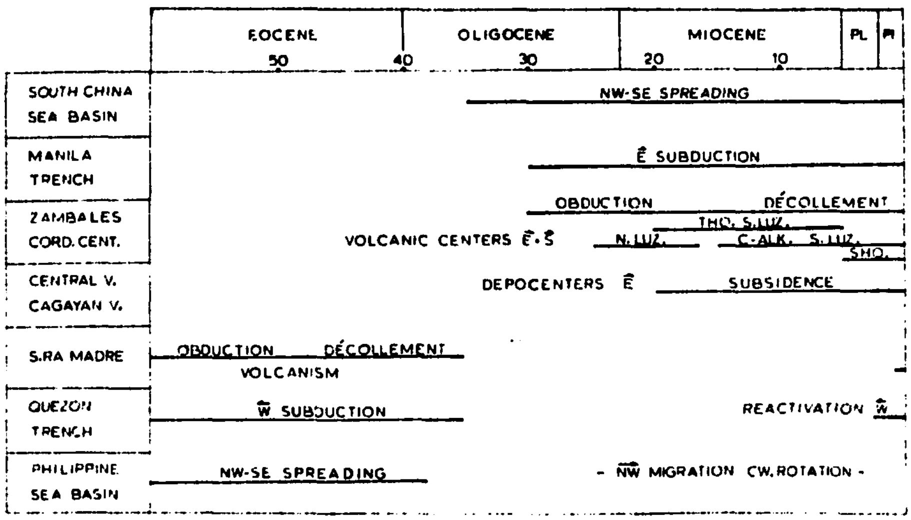


Figure 4. Summary of events in the island arc evolution of Northern Luzon from the Eocene to the present (from DeBoer et al., 1980)

The evolution of the Cagayan basin is summarized in Figure 5. Subsidence was initiated in the early Miocene and was concentrated in a north-south zone which is now the foothills of the Cordillera (Christian, 1964). Sedimentation began in the Oligocene with deposition of over 8,000 m of Oligocene and Miocene marine sediments (Durkee and Pederson, 1961). Proximal and distal turbidites compose about 80% of the marine sediments which also include shales, chalks, and biohermal limestones. The turbidites were derived from a west-northwest andesitic source, while biohermal limestones formed on the eastern shelf of the basin (Caagusan, 1980). Regional uplift of the area occurred in the Plio-Pleistocene (Christian, 1964) resulting in the deposition of 400 to 2,000 m of transitional marine and fluvial sediment of the Ilagan and Awidon Mesa formations (Figure 5) (Corby et al., 1951; Durkee and Pederson, 1961;

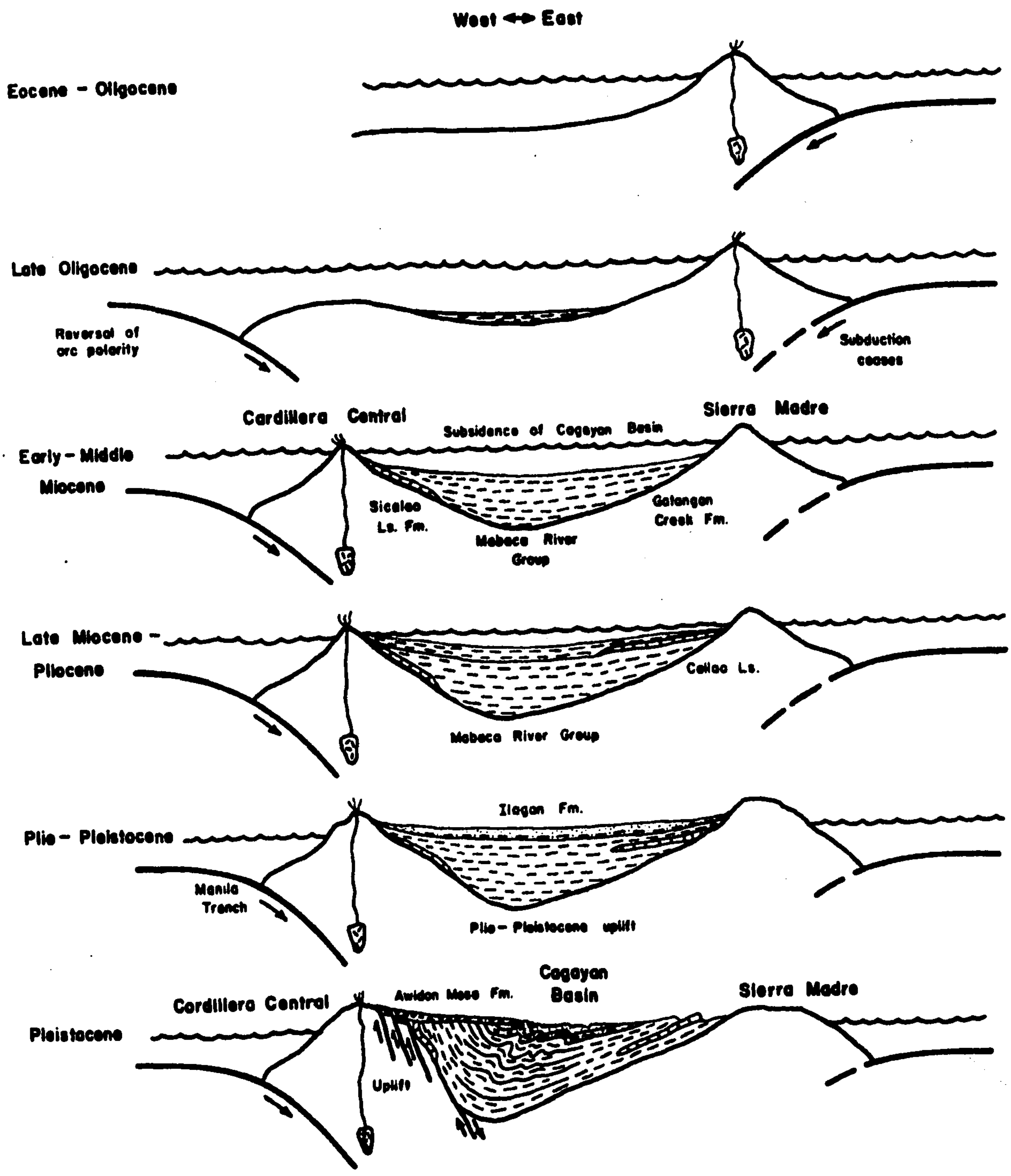


Figure 5. Schematic island arc evolution of the Cagayan basin

Tamesis, 1976). The Awidon Mesa Formation is a 400 m thick sequence of Pleistocene pyroclastic and fluvial sediments which conformably overlie the Ilagan Formation in the Cagayan Valley but unconformably overlie folded Miocene and Pliocene strata in the foothills of the Cordillera Central.

The sedimentary history of the Cagayan basin has been dominated by the active Cordillera Central volcanic arc (Durkee and Pederson, 1961). DeBoer et al. (1980) described the evolution of calc-alkaline volcanism in the Miocene to possible shoshonitic volcanism in the Pleistocene. As a result of volcanism and the corresponding geanticlinal uplift of the Cordillera Central, the basin became asymmetrical, and the axis migrated approximately 30 km eastward between the Miocene and Pleistocene (Figure 6) (Christian, 1964). In the middle to late Pleistocene, oversteepening of sediments in the Cordillera resulted in décollement and gravity sliding of the unstable uplifted sediments toward the basin (Christian, 1964; Caagusan, 1980). A series of asymmetric to overturned folds, the Cagayan anticlinal belt, were formed in the central part of the Cagayan Valley. Erosion of these folds has produced extensive outcrops of the Plio-Pleistocene Ilagan and Awidon Mesa formations, the main subject of this report.

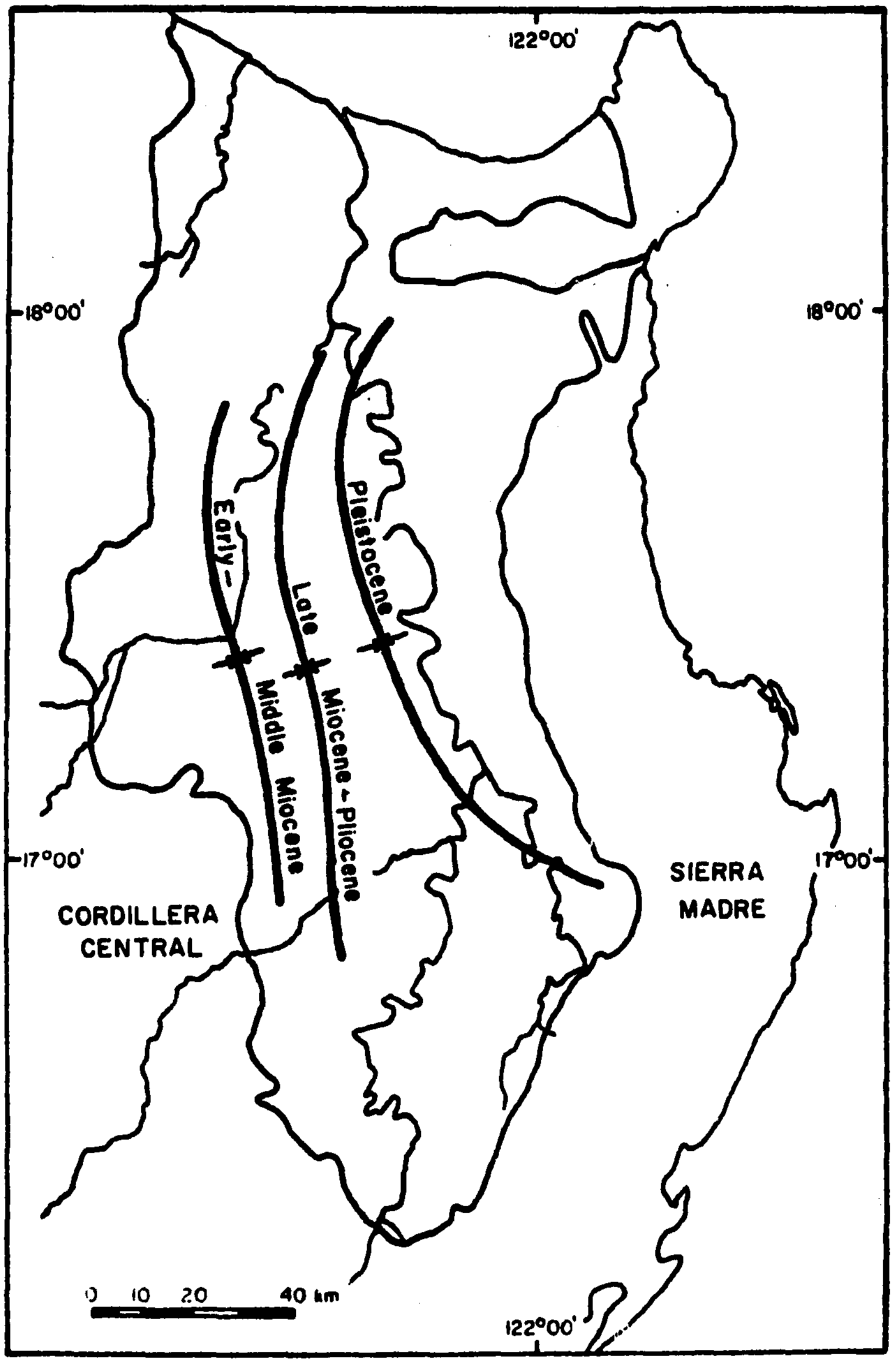


Figure 6. **Migration of Cagayan basin axis in Miocene through Pleistocene Epochs (after Caagusan, 1978)**

STRATIGRAPHY

Approximately 10,000 m of sediment has been deposited in the Cagayan basin since the Oligocene. The predominantly clastic succession was deposited on volcanic and metasedimentary rocks of the pre-Oligocene basement complex (Durkee and Pederson, 1961; Tamesis, 1976). The Oligocene and Miocene sediments, which were only briefly examined during reconnaissance, are of marine origin and are overlain by transitional marine and fluvial Plio-Pleistocene deposits.

As a result of a renewed oil exploration effort, the stratigraphic nomenclature for the basin has been recently revised (Tamesis, 1976). The revised nomenclature for the Oligocene and Miocene rocks contrasts significantly with that proposed by Corby et al. (1951), Kleinpell (1954), and Durkee and Pederson (1961) as indicated in Table 1. The nomenclature proposed by Durkee and Pederson (1961) will be used here because the nomenclature suggested by Tamesis (1976) has not been adequately defined for other workers to use. The Plio-Pleistocene stratigraphy will be described in detail following a brief review of the Miocene stratigraphy.

Miocene Stratigraphy

Sicalao Limestone

This formation is named after the Sicalao River on the western side of the valley (Durkee and Pederson, 1961). Five hundred forty-six meters of thin to massively bedded fossilferous carbonates were measured at the type section along Anaguan Creek, 3 km north-northwest of Rizal, Cagayan. The Sicalao Limestone can be traced nearly continuously along the western margin of the valley where it overlies mafic igneous rocks of the basement

Table 1. Stratigraphic nomenclature of the Cagayan Valley

Age		Corby et al., 1951	Kleinpell, 1954	Durkee and Pederson, 1961	Tamesis, 1976	Environments
Quaternary	Recent	Late Tertiary Volcanics; Ilagan Formation	Ilagan Formation	Awidon Mesa Formation	Awidon Mesa Formation	Fluvial
	Pleistocene	Late Tertiary Volcanics; Ilagan Formation; Magapit Ls.	Ilagan Formation	Awidon Mesa Formation	Awidon Mesa Formation	Fluvial
Tertiary	Pliocene	Late Tertiary Volcanics; Tuguegarao Sandstone	Ilagan Formation	Ilagan Formation	Ilagan Formation	Deltaic
	Miocene Upper	Tuguegarao Sandstone	Ilagan Formation	Mabaca River Group; Baliwag Formation	Cabagan Formation	Prodelta
	Miocene Middle	Callao Ls.	Callao Ls.	Mabaca River Group; Callao Ls.	Callao Ls.	Shelf to Deep Bathyal
	Miocene Lower	Lubuagan Coal Measures	Lubuagan Coal Measures	Mabaca River Group; Gatangan Creek Formation	Lubuagan Formation	Shelf to Deep Bathyal
	Oligocene	Ibulao Ls.	Ibulao Ls.	Sicalao Ls.	Dumata Formation	Supralittoral to Littoral
	Pre-Oligocene	Basement - Basic igneous volcanic rocks and metasediments				

complex. In some areas, however, the Sicalao Limestone is absent due to post-Sicalao faulting and erosion. The Formation was assigned an early Miocene age based on its orbitoid fauna (Durkee and Pederson, 1961).

Mabaca River Group

The Mabaca River Group, which is typically exposed along the Mabaca River, Kalinga-Apayao, includes all strata on the western side of the valley which overlie the Sicalao Limestone or basement and underlie the Pliocene Ilagan Formation (Durkee and Pederson, 1961). The group consists of three formations, the Asiga Formation, the Balbalan Sandstone, and Buluan Formation, which form a thick sequence of lutites, interbedded arenites, and some pyroclastic deposits. Based on reconnaissance observations, these deposits may be interpreted as shelf to deep bathyal clays and turbidites. These strata were previously referred to as the Lubuagan Formation by Corby et al. (1951) who did not designate a type area or section. The Lubuagan area is a structurally complicated region where it would be difficult to measure a complete stratigraphic section. The most complete section of the Mabaca River Group is 8,200 m thick and was measured 6 km east of Lubuagan between Toloctoc and Naneng. Based on the microfauna, the Mabaca River Group is considered to be of early to late Miocene or Miocene-Pliocene age (Durkee and Pederson, 1961). The Mabaca River Group is exposed along the western margin of the study area at the Chico River bridge and also at Pangul Anticline which is breached to the Buluan Formation.

Gatangan Creek Formation

On the east side of the Cagayan Valley, andesite flows of the basement complex are overlain by graywackes and claystones of the Gatangan Creek Formation (Durkee and Pederson, 1961). The Formation is named after Gatangan Creek, due east of Cabagan, Isabela. Here the Formation is 1,010 m thick and overlain by the Callao Limestone. The graywackes and claystones of the Gatangan Creek are here interpreted to be shelf and deep bathyal clay and turbidite deposits similar to those of the Mabaca River Group. Foraminiferal assemblages indicate that the Formation is early to middle Miocene in age, correlative with the lower part of the Mabaca River Group (Durkee and Pederson, 1961).

Callao Limestone

The Callao Limestone overlies the Gatangan Creek Formation on the east side of the valley. Corby et al. (1951) named the Callao Limestone and designated Barrio Callao, Cagayan as the type area. At Callao Canyon along the Pinacanaun de Tuguegarao River, the formation consists of 540 m of reef carbonates (Durkee and Pederson, 1961). The Callao was considered middle Miocene in age by Durkee and Pederson (1961) based on foraminifera collected at the type locality. The Philippine National Oil Company, on the basis of more complete paleontological evidence, has redated the Callao as late Miocene and Pliocene in age (E. V. Tamesis, PNOC, Manila, personal communication, 1978).

Baliwag Formation

Overlying the Callao Formation and underlying the Ilagan Formation is the Baliwag Formation, a valley forming claystone which was named and described by Vergara et al. (1959). It varies from 100 m thick along the eastern margin of the study area to 420 m thick at the type area along the Baliwag River near the Cagayan-Isabela provincial boundary. The Baliwag was considered to be late Miocene by Durkee and Pederson (1961) and correlative with the Buluan Claystones of the Mabaca River Group. Based on its stratigraphic occurrence, lithology, primary structures, and faunal content, the Baliwag is interpreted to be a prodelta deposit.

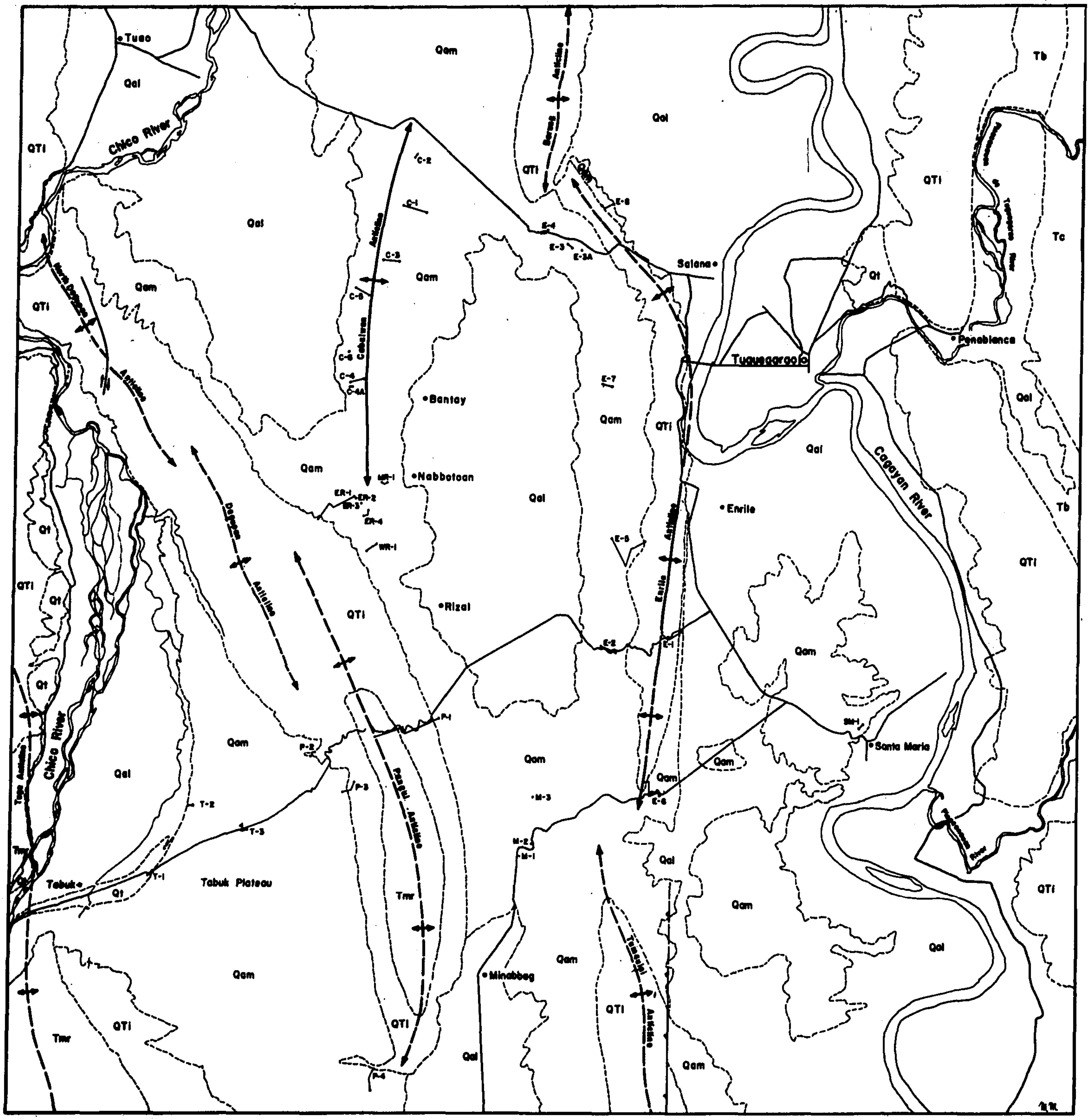
Plio-Pleistocene Stratigraphy

The Plio-Pleistocene deposits of the central Cagayan Valley consist of approximately 1,200 m of transitional marine and fluvial sediments. These deposits have largely been ignored by petroleum geologists due to their lack of oil potential. Both the Ilagan and Awidon Mesa formations were examined during this study because the contact relationship between the Pliocene and Pleistocene deposits had not been determined in the basin. The distribution of the Ilagan and Awidon Mesa formations, which are well-exposed along the flanks of eroded anticlines, is shown in Figure 7.

Ilagan Formation

The name Ilagan was proposed by Corby et al. (1951) for a sequence of poorly cemented sandstones 200 to 400 meters thick along the valley floor. No type area or type section was designated. It has generally been agreed by other workers that the name refers to outcrops of Plio-Pleistocene clastic sediments along the Ilagan River, south of Ilagan, Isabela Province, which exhibit the typical fluviatile depositional nature of the Formation (Kleinpell, 1954; Durkee and Pederson, 1961; Tamesis, 1976). No detailed description of the Formation has been previously given because

Figure 7. Geologic map of the central Cagayan Valley


Tuao
Qal
Qam
QTl
Chico River
Tb
Tc
Qt
Salana
Tuguegarao
Penablanca
Bantay
Nabbotoan
Enrile
Rizal
Cagayan River
Santa Maria
Tabuk
Tabuk Plateau
Tmr
Minabbag
Pasgal Anticline
Enrile Anticline
Dagupan Anticline
Cabalaran Anticline
Tamauini Anticline
Chico River
C-1
C-2
C-3
C-5
C-6
E-1
E-2
E-3
E-4
E-5
E-6
E-7
M-1
M-2
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WR-1
ER-1
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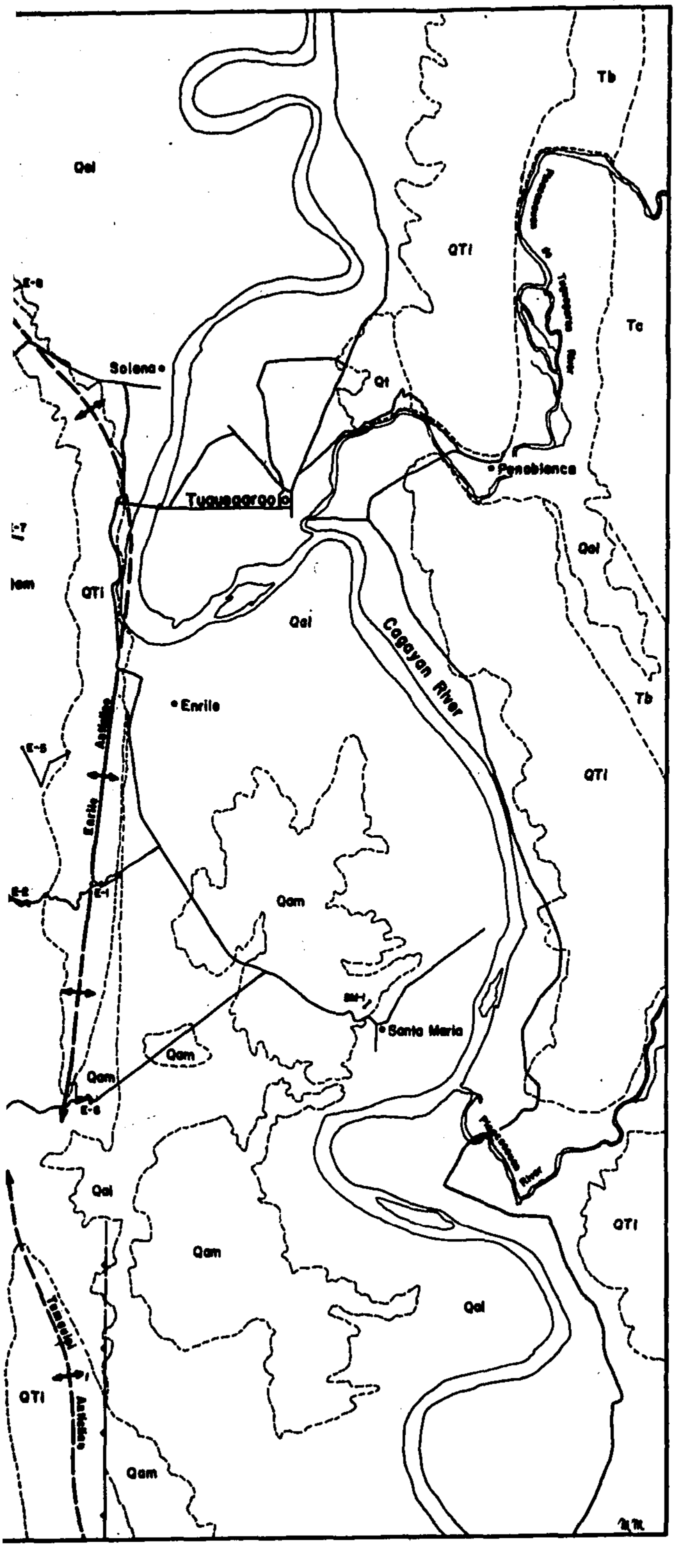


GEOLOGIC MAP of the CENTRAL CAGAYAN VALLEY, PHILIPPINES

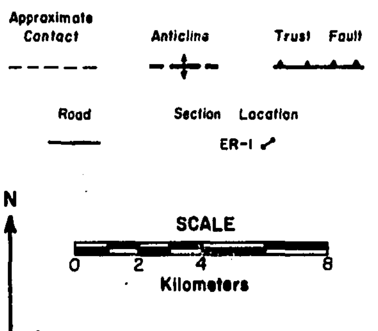
EXPLANATION

QUATERNARY

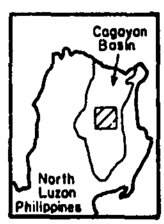
- Recent: Qal — Alluvium
- Pleistocene: Qt — Terrace
- Pleistocene: Qam — Awidon Mesa Formation
- Plio-Pleistocene: QTi — Ilagan Formation

TERTIARY

- Miocene: Tmr — Mabaca River Group
- Miocene: Tb — Baliwag Formation
- Miocene: Tc — Callao Limestone



Approximate mean declination varies from 0°14'W to 0°10'E



Map Location

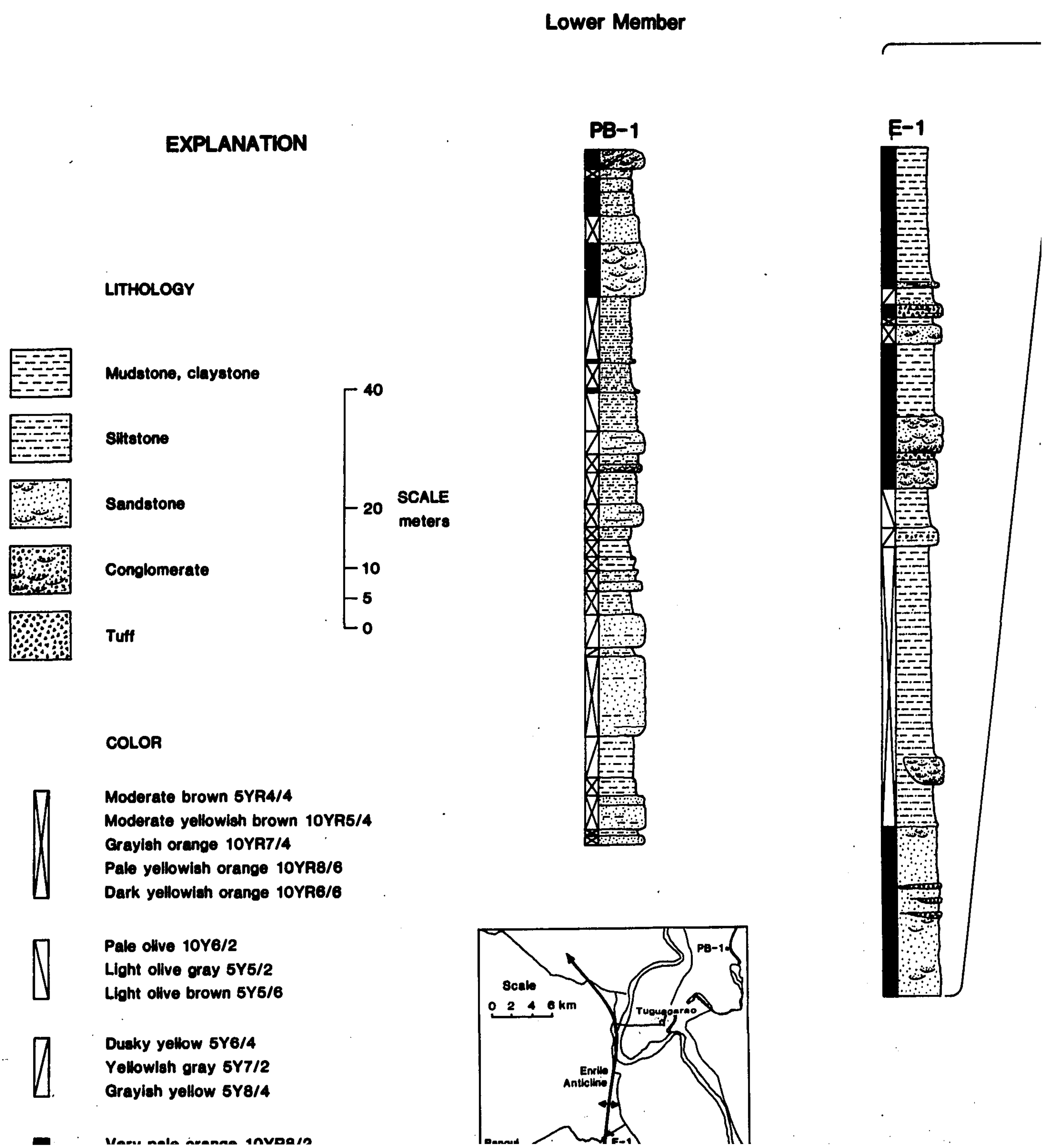
significant lateral lithological variations occur over short distances, and there was no economic incentive to study it.

The Ilagan Formation was studied in detail in the central Cagayan Valley to gain a better understanding of basin evolution and the contact relationship with the overlying Awidon Mesa Formation. In this area, the Formation, as mapped by Durkee and Pederson (1961), consists of a lower 150-310 m thick sequence of thin interbedded sandstones, siltstones, and mudstones, and an upper 500 m thick sequence of thicker, cross bedded conglomerates, sandstones, siltstones, and claystones. Based on a regional reconnaissance of the valley, Kleinpell (1954) suggested the Formation be divided into two members, a lower mudstone member, here called the Lower Member, and an upper sandstone, the Upper Member. Representative stratigraphic sections of both members are illustrated in Figure 8.

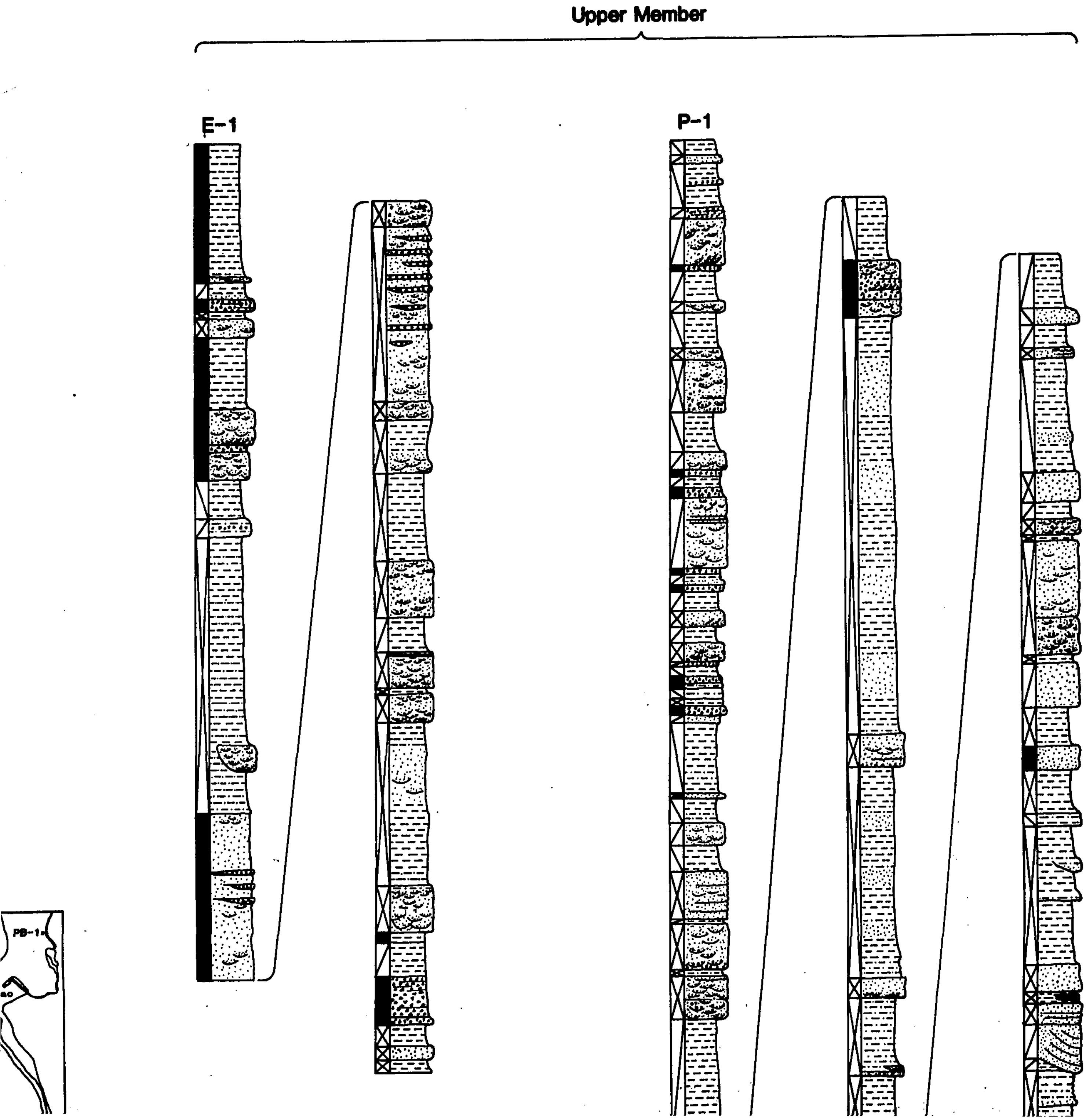
Lower Member The Lower Member is composed of thin, interbedded sandstones, siltstones, and mudstones (Figure 9a) which conformably overlie late Miocene and Pliocene claystones of the Baliwag Formation on the east side of the valley and the Mabaca River Group on the west side. The basal contact is gradational and mostly covered but distinct at the base of escarpments formed by the resistant Ilagan sandstones. The upper contact is the base of the first thick tabular body of trough cross bedded sandstone which fines upward to siltstone and a thick claystone. The dark greenish gray (5GY4/1) to pale yellowish orange (10YR8/6), fine to very fine grained sandstones and siltstones are laterally continuous, well-sorted, and usually exhibit parallel laminations, lenticular bedding or flaser bedding or small scale trough cross bedding. In the upper part of the Lower Member, more massive lenticular, medium to coarse grained

Figure 8. Graphic sections of the Ilagan Formation

GRAPHIC SECTIONS OF THE ILAG

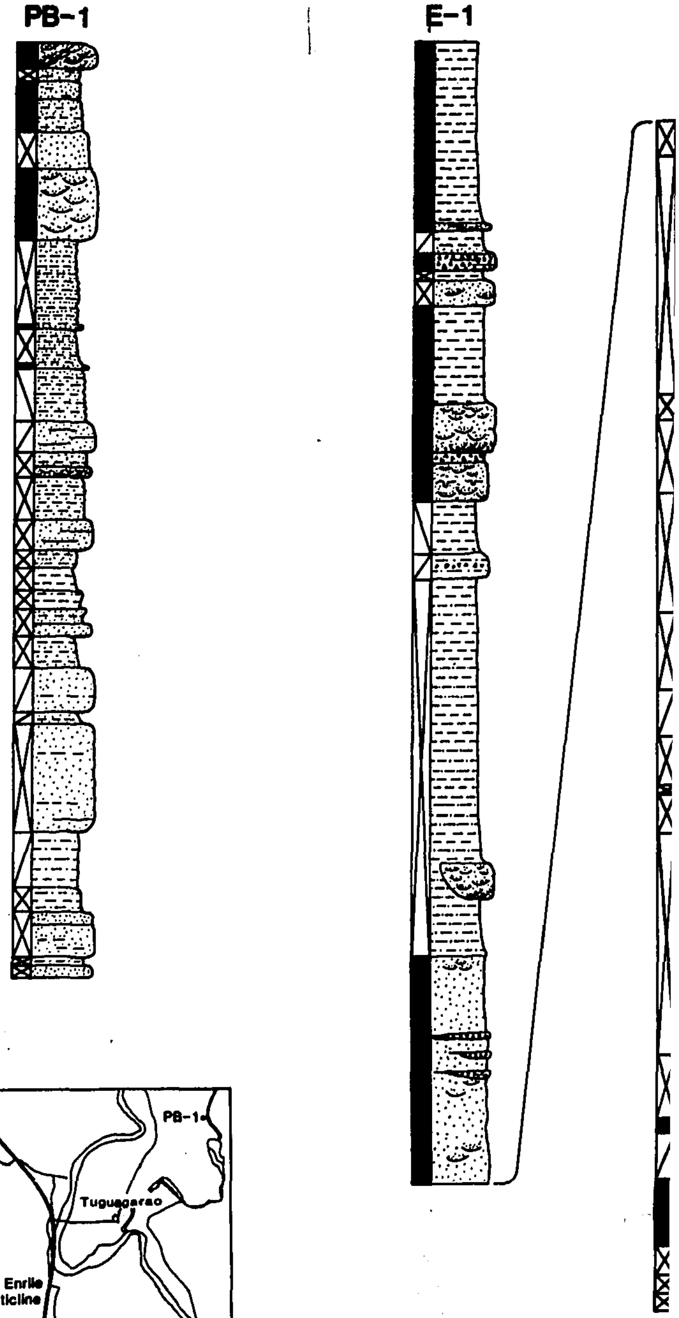


SECTIONS OF THE ILAGAN FORMATION



GRAPHIC SECTIONS OF THE ILAGAN

Lower Member



EXPLANATION

LITHOLOGY

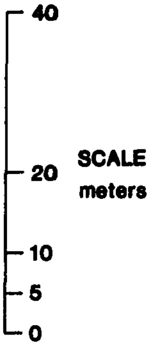
Mudstone, claystone

Siltstone

Sandstone

Conglomerate

Tuff



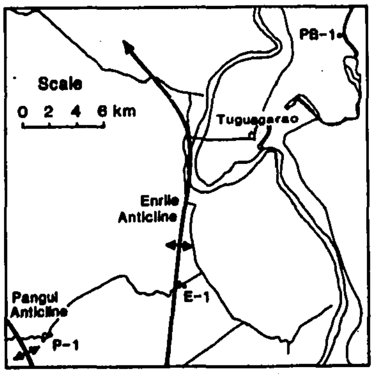
COLOR

Moderate brown 5YR4/4
Moderate yellowish brown 10YR5/4
Grayish orange 10YR7/4
Pale yellowish orange 10YR8/6
Dark yellowish orange 10YR6/6

Pale olive 10Y6/2
Light olive gray 5Y5/2
Light olive brown 5Y5/6

Dusky yellow 5Y6/4
Yellowish gray 5Y7/2
Grayish yellow 5Y8/4

Very pale orange 10YR8/2
Medium light gray N6
Very light gray N8
White N9



LOCATION OF SECTIONS

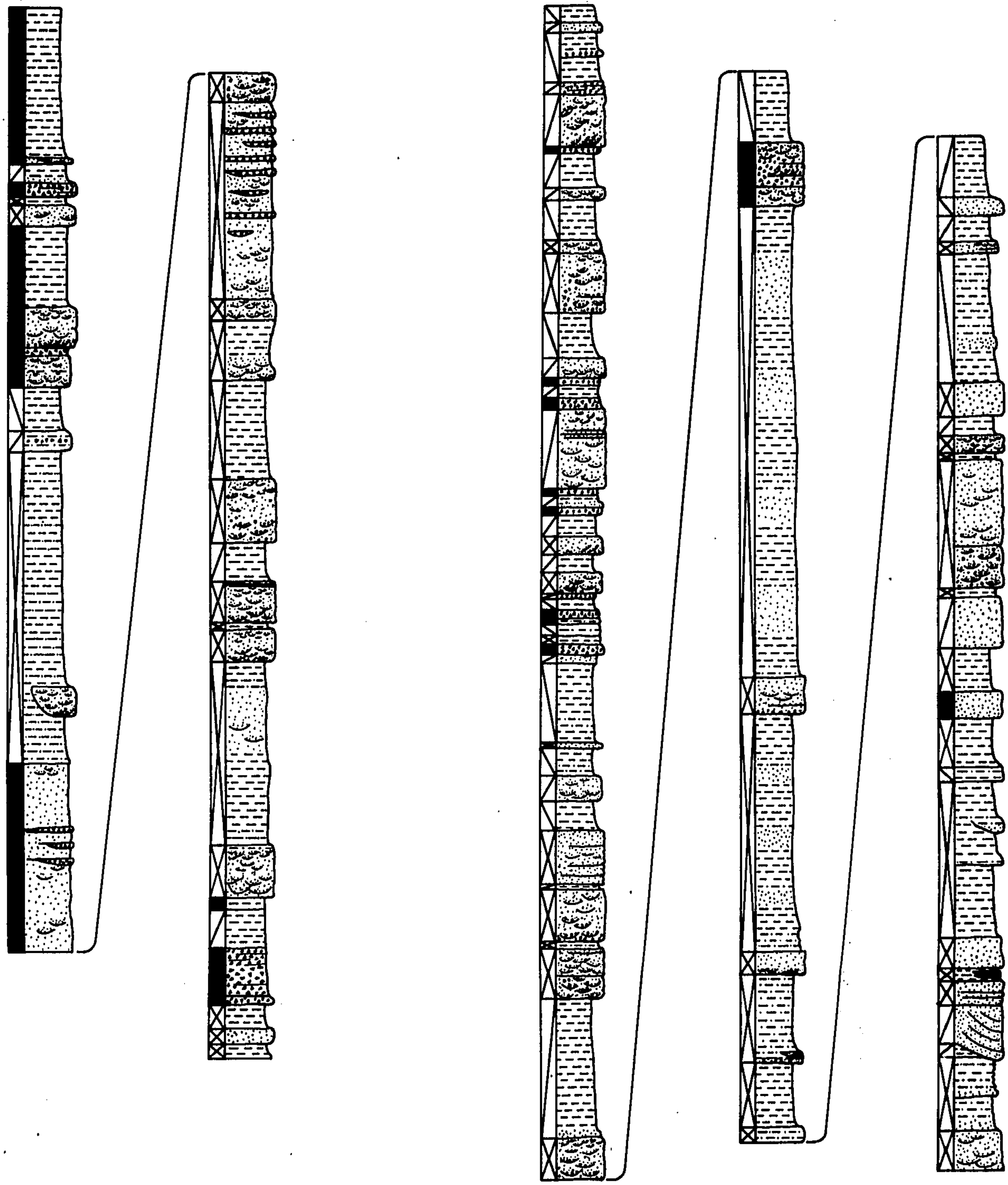
C SECTIONS OF THE ILAGAN FORMATION

r

Upper Member

E-1

P-1



ONS

Figure 9. Photographs of the Ilagan and Awidon Mesa formations

a. Lower Member of the Ilagan Formation, interbedded sandstones, siltstones, and mudstones, along the Pinacanauan de Tuguegarao River

b. Ilagan Formation Upper Member, conglomerates and sandstones which fine upwards to siltstones and claystones; east flank, Pangul Anticline

c. Awidon Mesa, type area of the Awidon Mesa Formation, formed by 300 m of valley filling pyroclastic deposits 6 km northwest of Lubuagan, Kalinga-Apayao Province

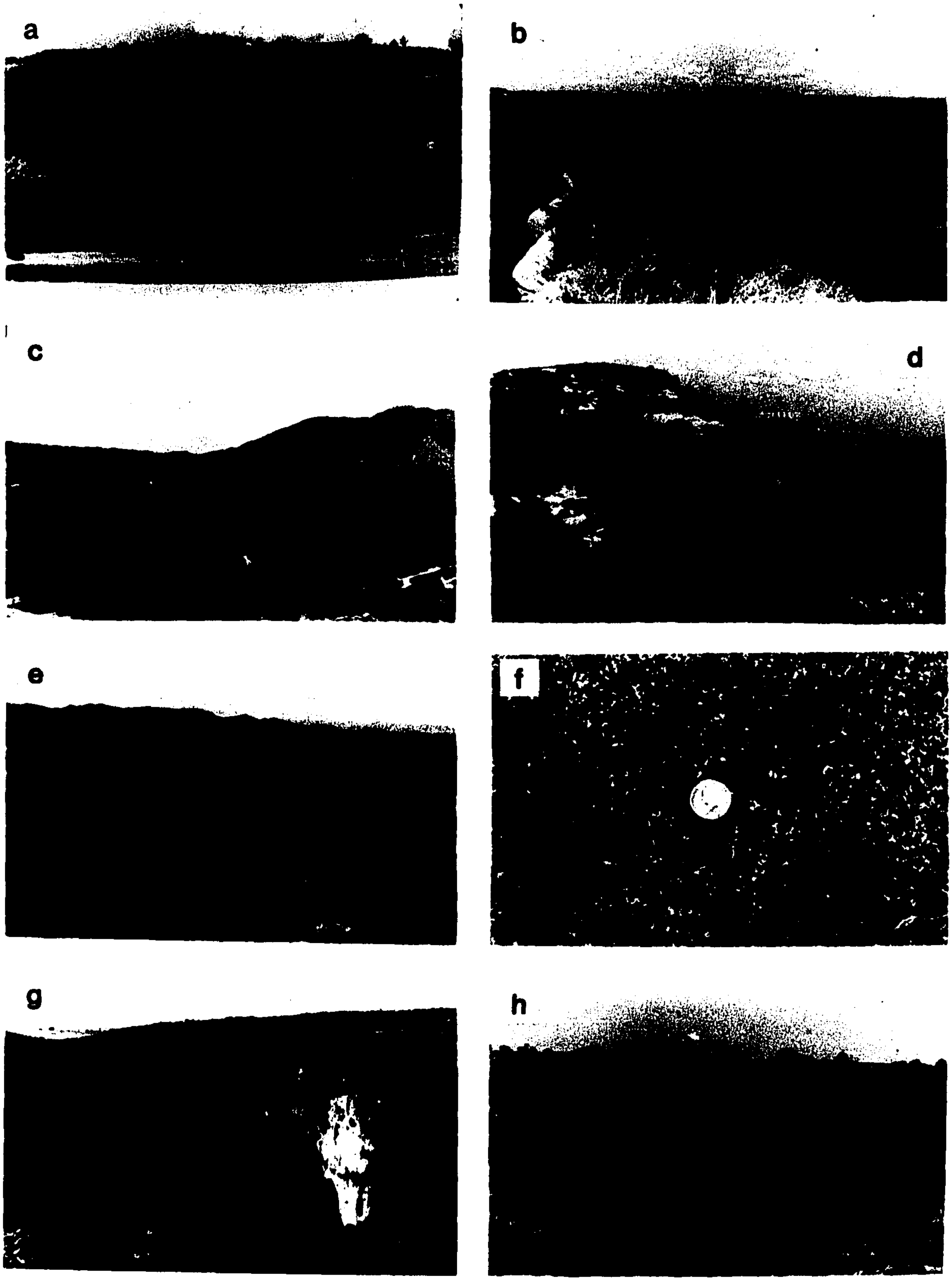
d. Awidon Mesa Formation, fluvial sandstones, siltstones, and claystones at the Espinosa Ranch, Pangul Anticline

e. Angular unconformity formed by the flat-lying Awidon Mesa Formation and folded Miocene Mabaca River Group in Kalinga foothills near the Chico River

f. Bipyramidal quartz granules derived from the Awidon Mesa Formation which form an erosional residue where the formation is present

g. Liwan Pyroclastic Complex, west flank of Enrile Anticline near village of Liwan, Kalinga-Apayao

h. Tabuk Pyroclastics capping the Tabuk plateau along Aliog Creek


a
b
c
d
e
f
g
h

sandstones up to 9 m thick occur. The bedding is often indistinct, but small and large scale trough cross bed sets up to 50 cm thick are common. Planar cross beds up to 30 cm thick were also present at one locality. Blackish red (5R2/2) calcareous and ferruginous disk to blade shaped concretions up to boulder size are present along numerous horizons within the Lower Member. Gastropods and pelecypods are locally abundant in sandstones while plant fragments and burrows are common throughout the member. Numerous shark teeth found on the surface appear to have been eroded from the Lower Member.

Upper Member Plio-Pleistocene conglomerates, sandstones, siltstones, and claystones that occur between the first tabular sandstone body which fines upward to claystone and the first tuff-breccia deposit or quartz granule-bearing conglomerate form the Upper Member of the Ilagan Formation. The Upper Member outcrops along the flanks of Enrile and Pangul anticlines in the central Cagayan Valley. Graphic sections of the Upper Member at these anticlines are illustrated in Figure 8. The entire Upper Member is exposed at Pangul Anticline where it attains a thickness of 500 m (Figure 9b). Only the upper part of the Member is exposed at Enrile Anticline. The conglomerates, sandstones, and claystones form laterally extensive sheet-like deposits which are commonly traceable for at least several kilometers.

The sediments of the Upper Member typically occur in fining upward sequences. In the lower part of the Member, poorly sorted, medium to coarse grained, pale yellowish orange (10YR8/6) sandstones grade upward to moderately sorted, fine grained sandstones and siltstones which are overlain by massive, pale olive (10Y6/2) claystones which are often sandy.

There is also a coarsening upward throughout the Member as polymictic granule to pebble conglomerates become more common and thicker in the upper part of the Member. The polymictic conglomerates, which are dominantly clast supported, are composed primarily of porphyritic andesite, basalt, metasedimentary clasts, and chert. Primary structures characteristically grade upward from large scale trough cross bed sets up to 1 m thick in the conglomerates and coarse sandstones to small scale trough cross beds in the finer sandstones. Heavy mineral-rich layers and grain size variations accent the cross bedding. Climbing ripple lamination may also occur in the finer sandstones. Scour-and-fill is common in the conglomerates and sandstones while load structure, convolute bedding, and calcareous concretions occur occasionally. The siltstones are usually thin and characterized by parallel lamination. The overlying claystones are massive, commonly reaching 20 m in thickness. The sandstones commonly are thick, reaching 12 m while the conglomerates only thicken in the upper portion of the member. The fining upward sequences commonly attain a thickness of 20 m.

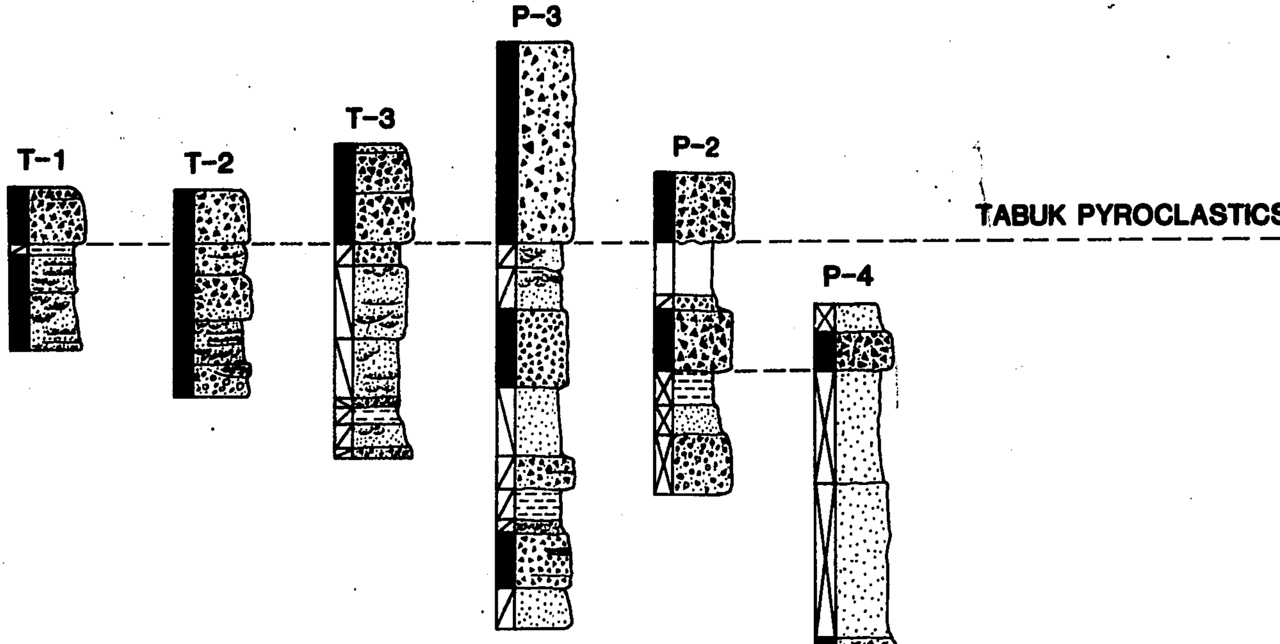
Thin, lenticular, very light gray (N8) to very pale orange (10YR8/2) tuffs occur throughout the Upper Member associated with the upper sandstones, siltstones, and claystones of the fining upward sequences. Along the northwest flank of the Pangul Anticline, the tuffs occur in three distinct tuffaceous intervals which commonly contain permineralized logs and abundant leaf impressions, primarily ferns. Trace fossils such as burrows and root casts occur in the tuffs as well as in the associated siltstones and claystones. No megafossils have been found in the Upper Member of the Ilagan.

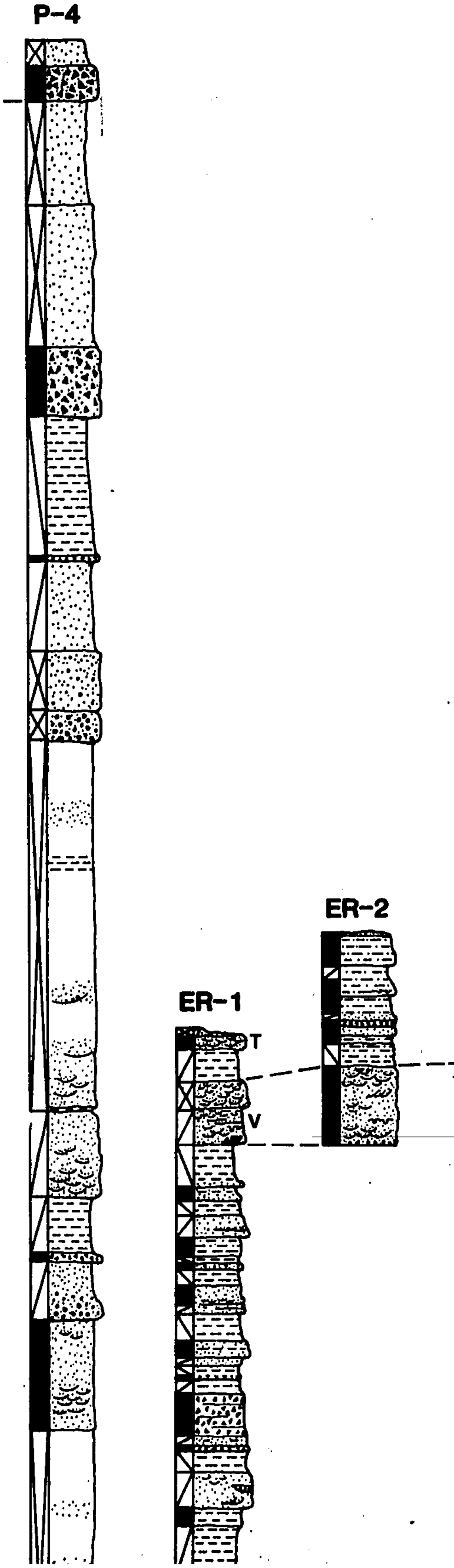
Awidon Mesa Formation

The Awidon Mesa Formation was named and described by Durkee and Pederson (1961). They used the term to describe middle Pleistocene tuffaceous sediments of a dacitic type which are characterized by the presence of bipyramidal quartz (approximately 8%), euhedra of hornblende, and sodic feldspar. The Formation unconformably overlies folded strata of Miocene age at its type locality, Awidon Mesa, in the Cordillera Central near Lubuagan, Kalinga-Apayao Province. It attains a thickness of 300 m at Awidon Mesa and discontinuously extends out into the Cagayan Valley, where it overlies older tuffaceous sediments of the Ilagan Formation (Durkee and Pederson, 1961). The contact relationship between the Awidon Mesa and the Ilagan Formation in the valley has not been previously studied. One of the major objectives of this study was to determine the character and distribution of the Awidon Mesa Formation in the Cagayan Valley.

The Formation grades from a thick, massive, valley fill sequence of dacitic pyroclastic deposits, primarily tuff-breccias, at the type locality in the mountains (Figure 9c) to a sequence of thinner tuff-breccias and tuffs that are interbedded with tuffaceous fluvial sediments (Figure 9d) in the central Cagayan Valley. The Formation is well-exposed in the valley along the Tabuk plateau, Pangul, Cabalwan, and Enrile anticlines (Figure 7) where most stratigraphic sections were measured (Figure 10 and 11). In this area, the formation reaches a maximum thickness of 400 m at southern Pangul Anticline and thins to the east and north. The basal contact of the Awidon Mesa Formation grades laterally from an unconformable relationship with the underlying folded Miocene sediments in the mountains (Figure 9e) to a conformable relationship with the Upper Member of the Ilagan in the

Figure 10. Graphic sections of the Awidon Mesa Formation, Tabuk plateau, and Pangul Anticline





GRAPHIC SECTIONS

OF THE

AWIDON MESA FORMATION, TABUK PLATEAU AND PANGUL ANTICLINE

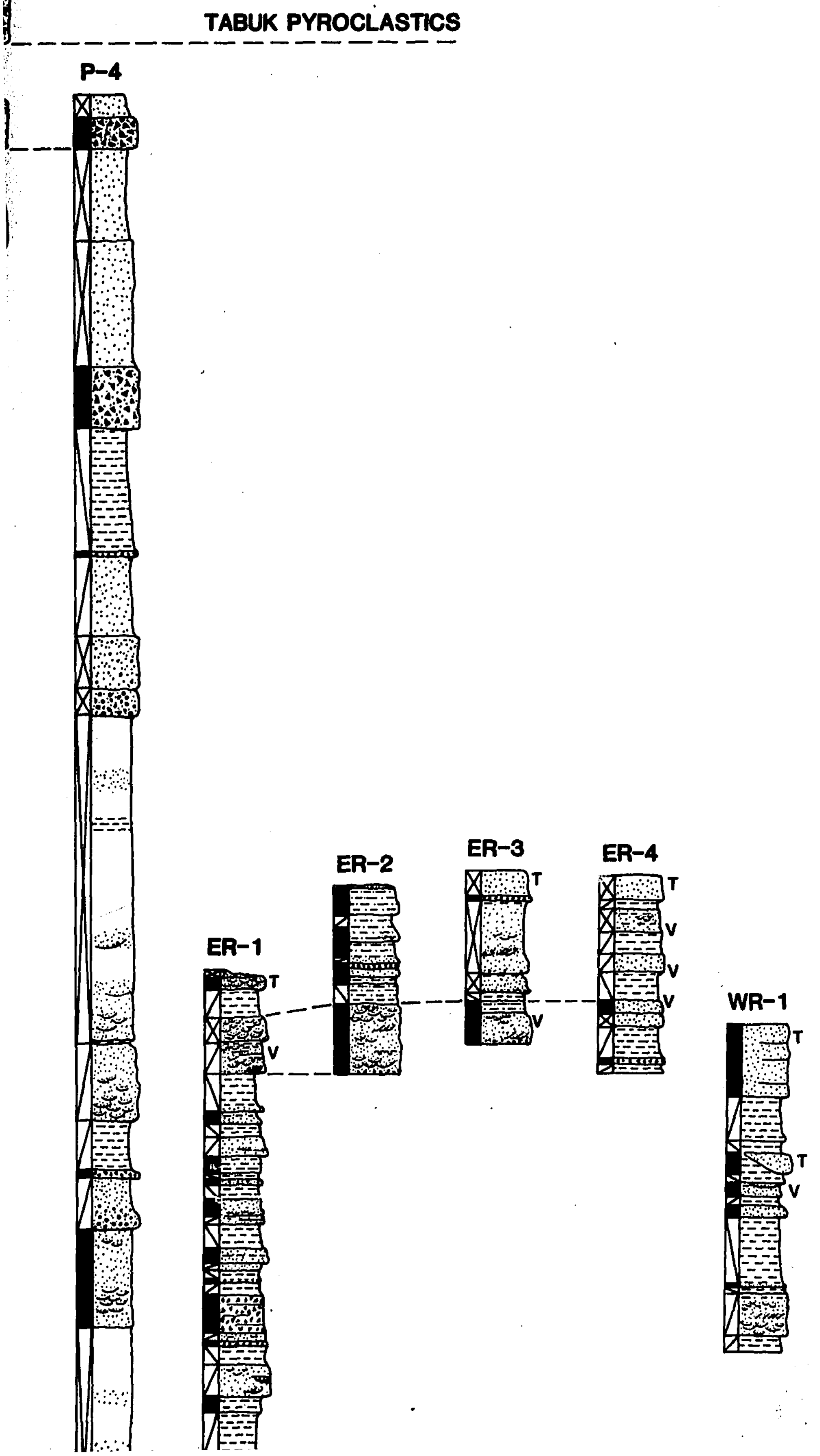
EXPLANATION

LITHOLOGY

Claystone

COLOR

Moderate brown 5YR4/4
Moderate yellowish brown 10YR5/4
Dark yellowish orange 10YR6/6
Grayish orange 10YR7/4


TABUK PYROCLASTICS
P-4
ER-1
ER-2
ER-3
ER-4
WR-1
T
V

EXPLANATION

LITHOLOGY

- Claystone
- Siltstone
- Sandstone
- Conglomerate
- Tuff-breccia
- Tuff
- Covered

COLOR

- Moderate brown 5YR4/4
- Moderate yellowish brown 10YR5/4
- Dark yellowish orange 10YR6/6
- Grayish orange 10YR7/4

- Pale olive 10YR6/2
- Light olive gray 5Y5/2
- Light olive brown 5Y5/6

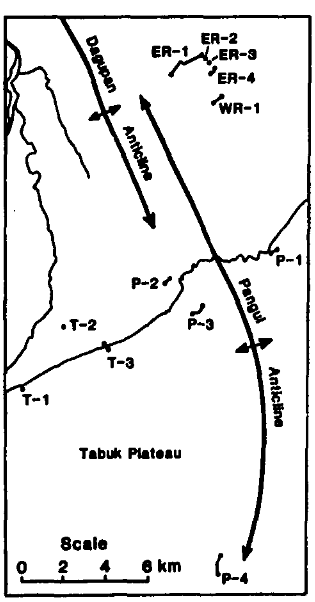
- Dusky yellow 5Y6/4
- Grayish yellow 5Y8/4
- Yellowish gray 5Y7/2

- White N9
- Very light gray N8
- Light gray N7
- Very pale orange 10YR8/2

V
Vertebrate fossils, surface

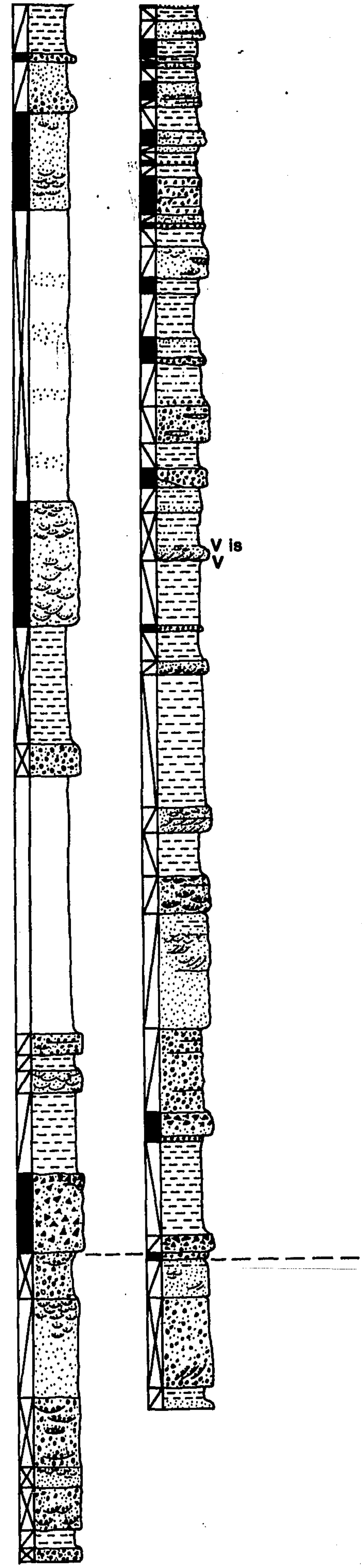
V is
Vertebrate fossils, in situ

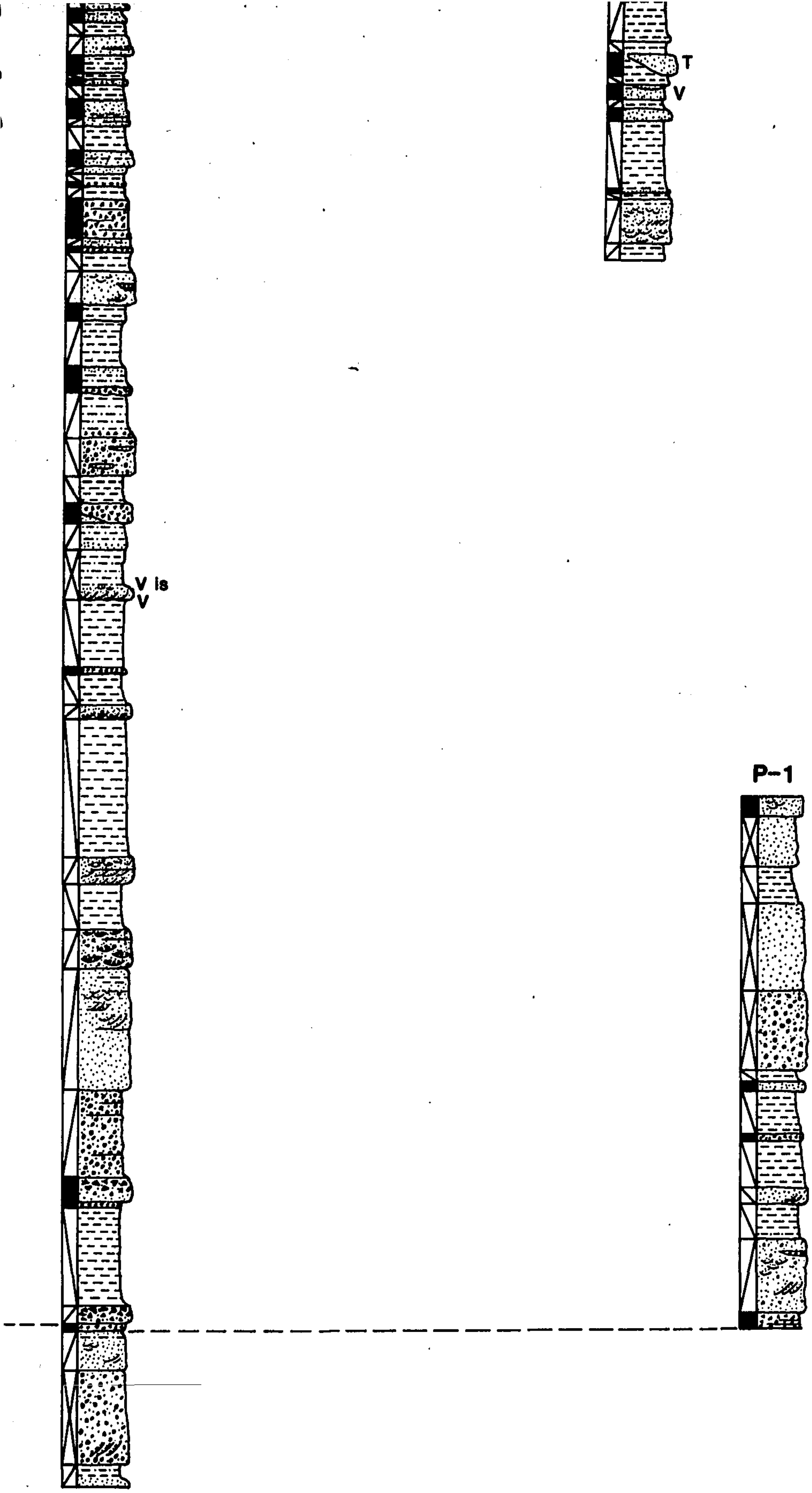
T
Tektite locality



LOCATION OF SECTIONS

SCALE meters: 0, 5, 10, 20, 40





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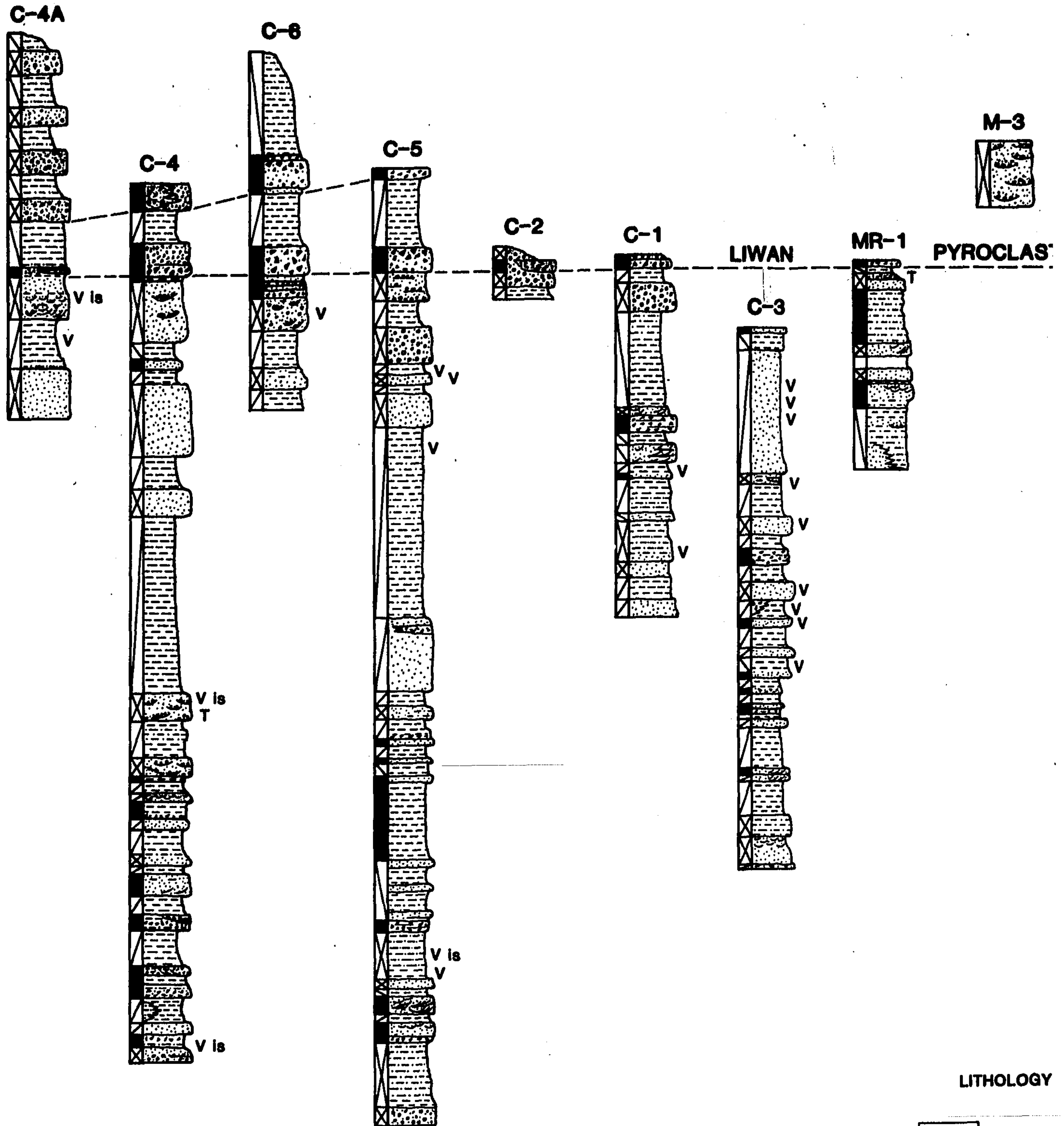
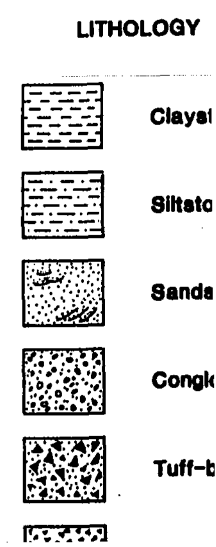
V
P–1

Figure 11. Graphic sections of the Awidon Mesa Formation, Cabalwan and Enrile anticlines

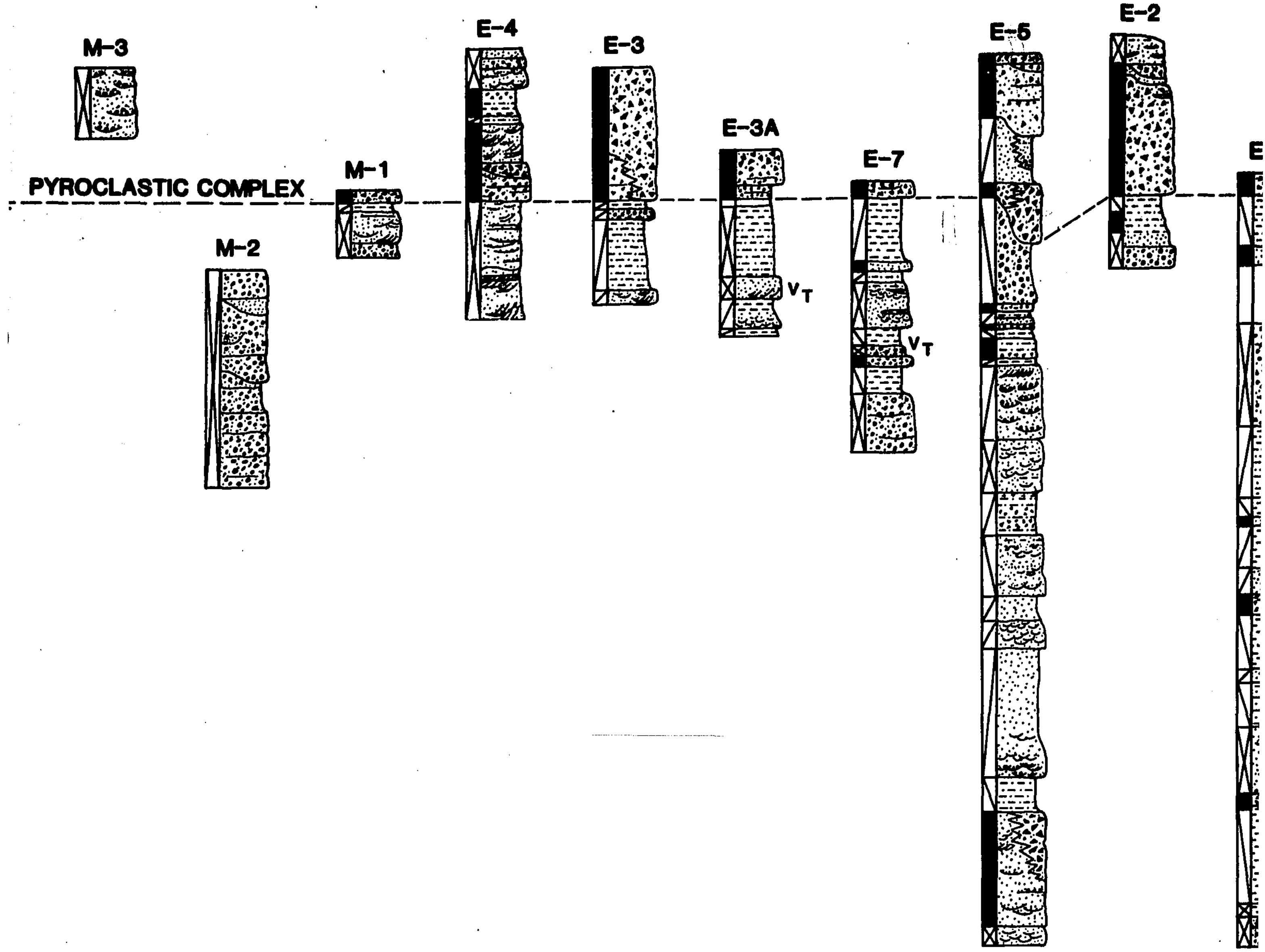




GRAPHIC SECTIONS

OF THE

AWIDON MESA FORMATION,



EXPLANATION

LITHOLOGY

Claystone

Siltstone

Sandstone

Conglomerate

Tuff-breccia

COLOR

Moderate brown 5YR4/4
Moderate yellowish brown 10YR5/4
Grayish orange 10YR7/4
Dark yellowish orange 10YR6/6

Light olive brown 5Y5/6
Pale olive 10Y6/2
Light olive gray 5Y5/2

Dusky yellow 5Y6/4
Grayish yellow 5Y8/4
Yellowish gray 5Y7/2

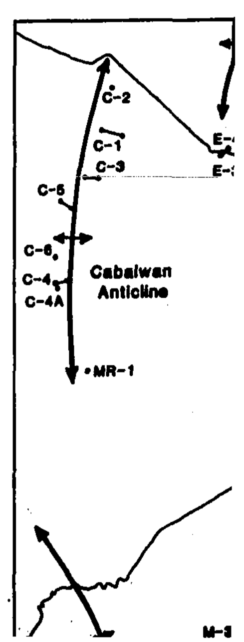
V
Vertebrate fossils, surface

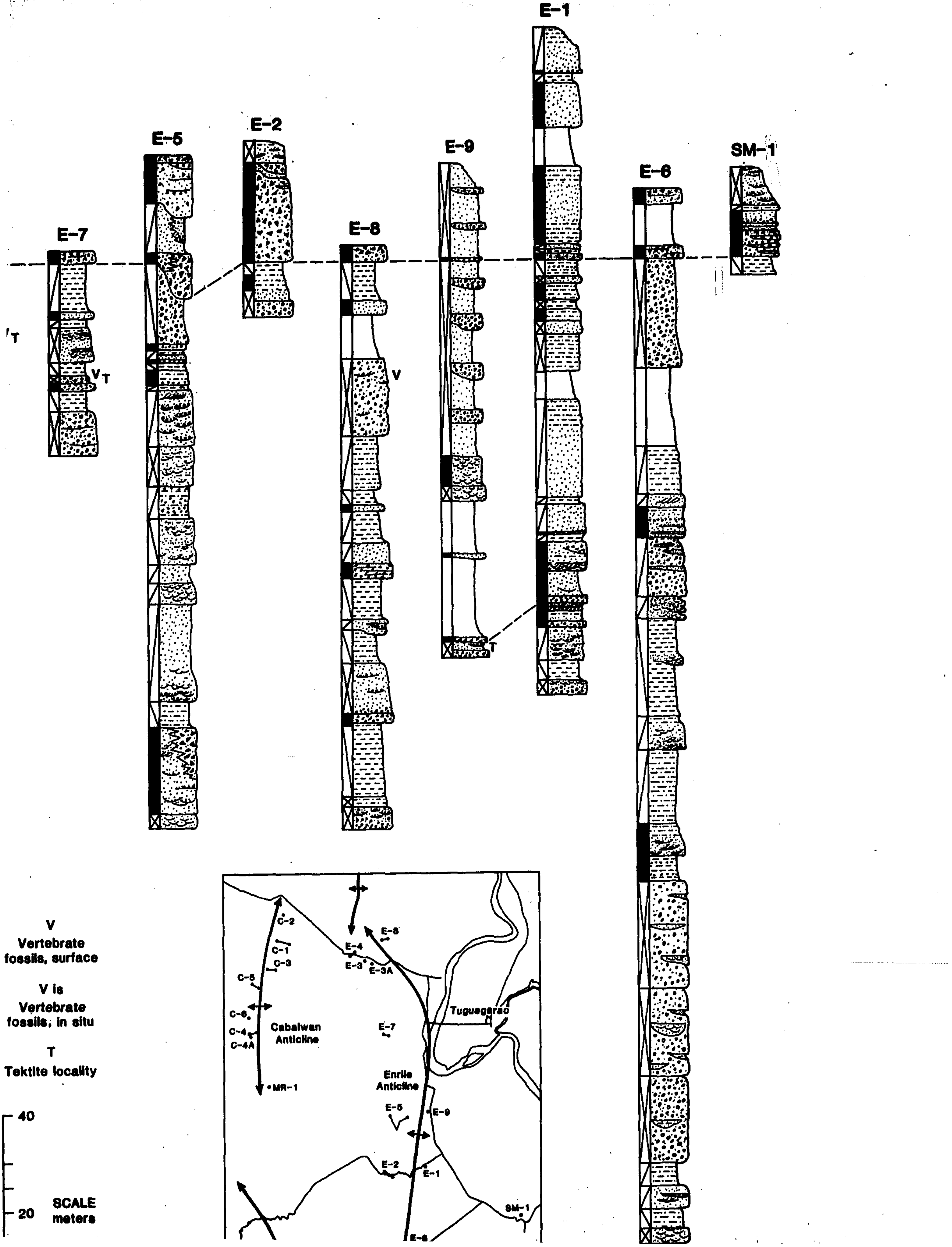
V is
Vertebrate fossils, in situ

T
Tektite locality

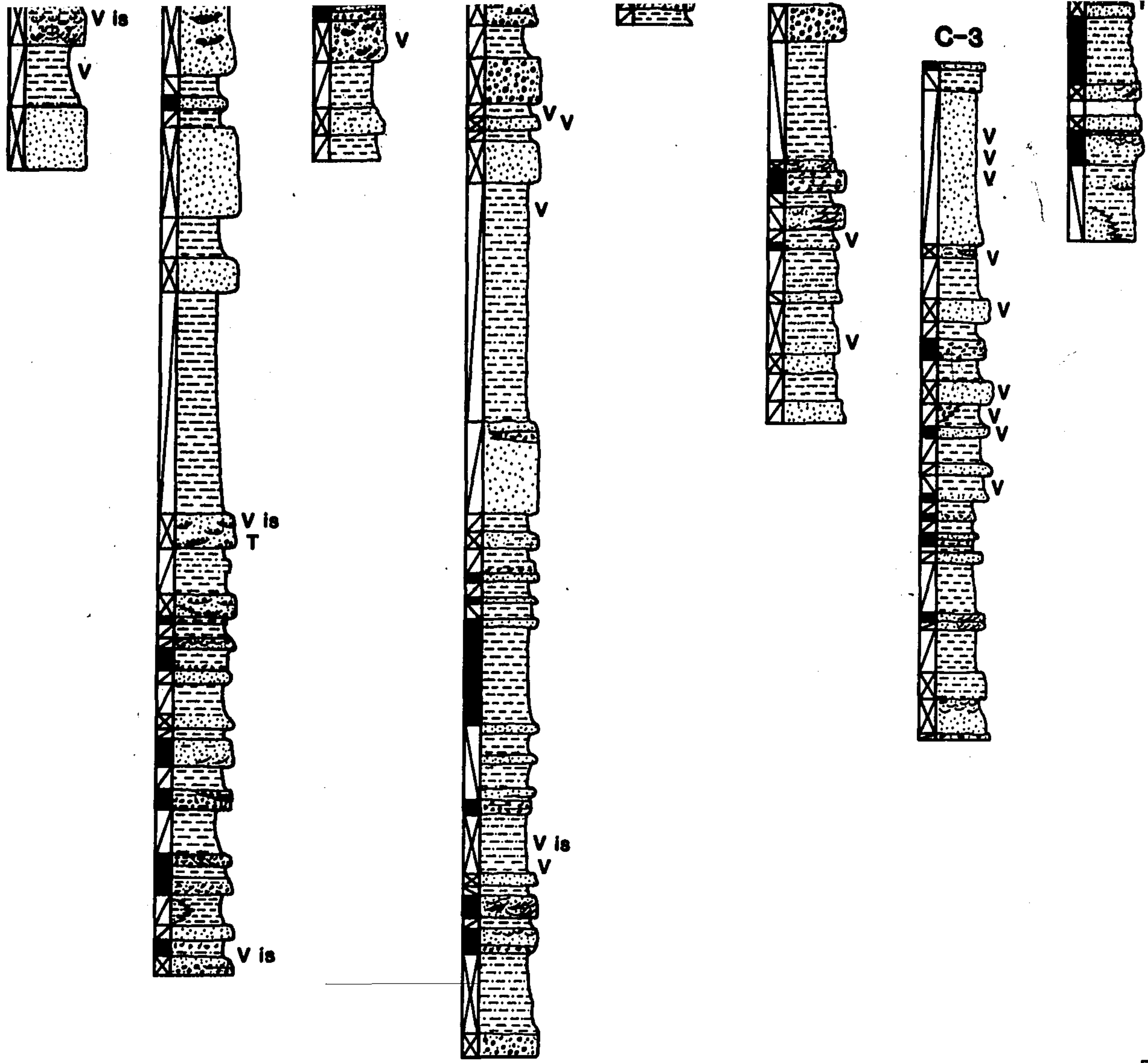
40

20 SCALE meters





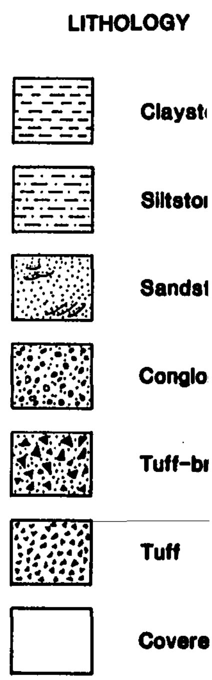
E-7
E-5
E-2
E-8
E-9
E-1
E-6
SM-1
T
V T
V
T
V
Vertebrate fossils, surface
V is
Vertebrate fossils, in situ
T
Tektite locality
40
20
SCALE
meters
C-2
C-1
C-3
C-5
C-6
C-4
C-4A
Cabalwan Anticline
MR-1
E-8
E-4
E-3
E-3A
E-7
Tuguegarao
Enrile Anticline
E-5
E-9
E-2
E-1
SM-1
E-6

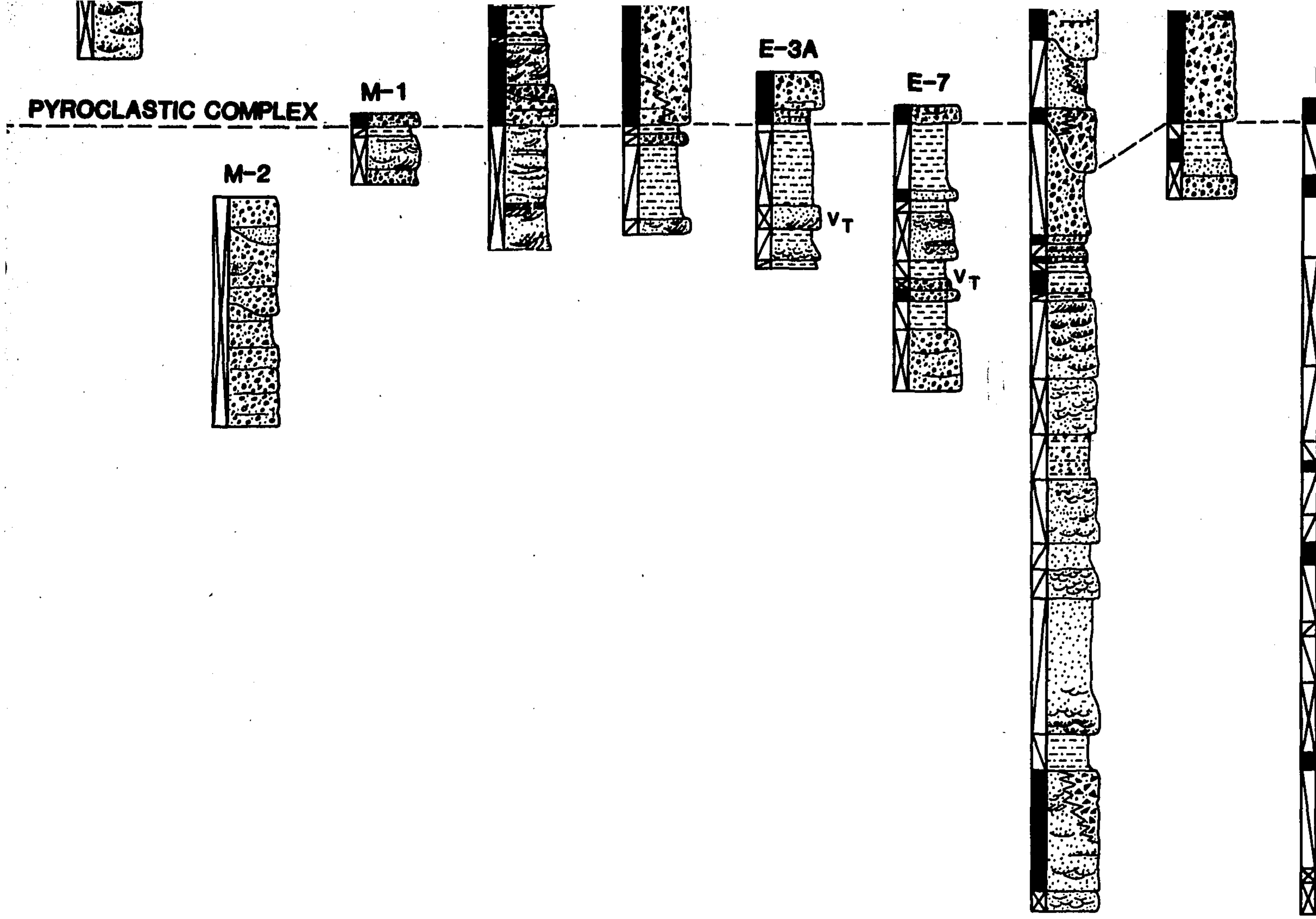


GRAPHIC SECTIONS

OF THE

AWIDON MESA FORMATION, CABALWAN AND ENRILE ANTICLINES, CENTRAL CAGAYAN VALLEY





EXPLANATION

LITHOLOGY

- Claystone
- Siltstone
- Sandstone
- Conglomerate
- Tuff-breccia
- Tuff
- Covered

COLOR

Moderate brown 5YR4/4
Moderate yellowish brown 10YR5/4
Grayish orange 10YR7/4
Dark yellowish orange 10YR6/6

Light olive brown 5Y5/6
Pale olive 10Y6/2
Light olive gray 5Y5/2

Dusky yellow 5Y6/4
Grayish yellow 5Y8/4
Yellowish gray 5Y7/2

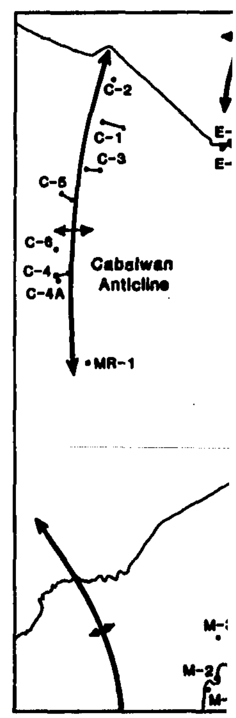
Very light gray N8
Very pale orange 10YR8/2
White N9

V
Vertebrate fossils, surface

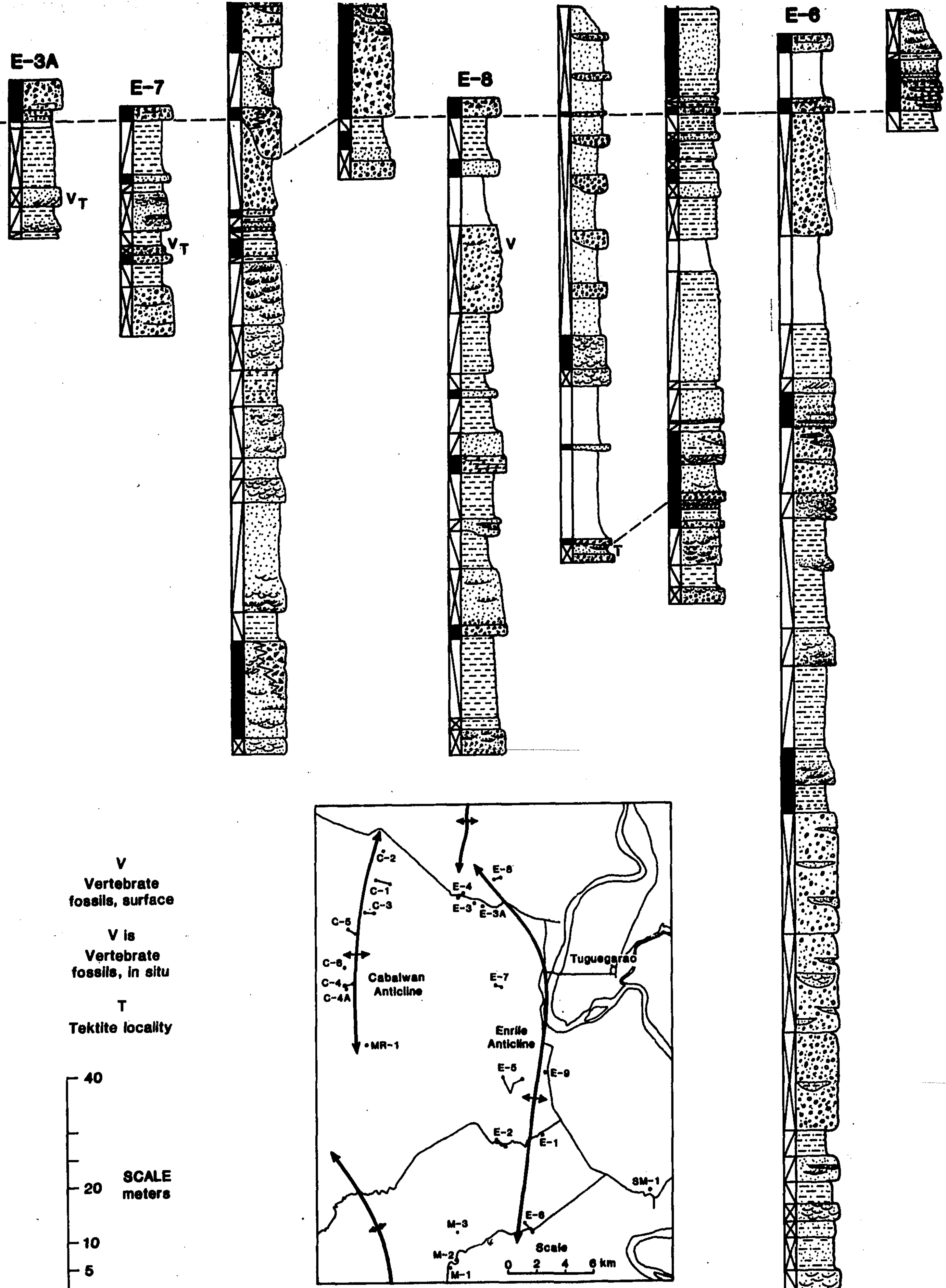
V is
Vertebrate fossils, in situ

T
Tektite locality

40
20
10
5
0
SCALE meters



LOCATI



LOCATION OF SECTIONS

5/4

Cagayan Valley. The base of the oldest dacitic tuff-breccia deposit or the first quartz granule bearing conglomerate has been designated as the contact between the Awidon Mesa Formation and the underlying Ilagan in the valley (Vondra et al., 1981).

The textures of the pyroclastic deposits vary considerably from very well-sorted, white (N9) tuffs to tuff-breccias. The tuff-breccias are formed by subangular or rounded lapilli and blocks in a poorly sorted, light gray (N7) dacitic tuff matrix. In some of the younger deposits in the valley, blocks several meters in diameter occur, the largest having a long axis of 5.5 meters. The pyroclasts in the coarser pyroclastic deposits are dominantly matrix supported. Orientation varies as pyroclasts in some deposits are oriented parallel to the bedding while pyroclasts in other deposits have random orientations. The pyroclasts weather in relief giving each deposit a characteristically knobby, grayish red (5R4/2) surface.

The composition of the tuff breccia matrix is uniform throughout the study area, but the dominant lithology of the pyroclasts varies. All the deposits have a light gray (N7) moderately indurated dacitic tuff matrix with a low clay content and a minor but significant amount of bipyramidal quartz. The quartz commonly forms a sparkling erosional residue on the surface where the Awidon Mesa Formation is present (Figure 9f). Pyroclast lithology varies with the age and location of the pyroclastic deposits. The youngest deposits which form the Tabuk Plateau at the base of the Cordillera Central contain primarily subangular, equant, very light gray (N8) dacite and light gray (N7) andesite pyroclasts. Some light greenish gray (5GY8/1) and pale red (5R6/2) andesite-dacite pyroclasts and minor amounts of basalt also occur. The older pyroclastic deposits exposed to the east along the flanks of Cabalwan, Pangul, and Enrile anticlines contain primarily rounded, equant to disk shaped pumice pyroclasts. Minor amounts of subrounded basalt pebbles are also sometimes present.

A variety of sedimentary structures occur in the pyroclastic deposits. The basal contacts of the tuff-breccias are usually sharp with little or no relief. Scour and fill structures, however, are present at the base of some deposits. The tuff-breccia deposits are characteristically massive but often exhibit graded bedding or reverse graded bedding. Gas escape structures (fumaroles), vertically oriented pipe-like features which lack a finer silt-clay matrix, occur in many of these deposits. In most outcrops, several flow units may be distinguished separated by sharp contacts. Small and large scale low angle trough cross-bedding often occurs at the base or top of these units. Some massive beds or beds with normal or reverse graded bedding grade to cross bedded deposits laterally indicating reworking by streams. The tuffs also commonly appear massive but often display faint small scale cross bedding or climbing ripple laminations indicating reworking by water or wind. Some deposits are stratified indicating they are probably primary fall deposits. Mantle bedding, a diagnostic feature of fall deposits (Sparks and Walker, 1973), was not observed, however, due to the limited lateral extent of the tuffs in outcrop.

Outcrop characteristics of the pyroclastic rocks were carefully examined to provide a basis for the local correlation of strata and possible subdivision of the Awidon Mesa Formation for the documentation of archaeological sites or fossil localities. The tuffs are usually lenticular and thin but may attain a thickness as great as 2 m. They are difficult to use for correlation, however, as they usually pinch out over a short distance or are covered. The tuff breccias are more valuable for correlation because they occur as thicker tabular deposits that are traceable for up to 15 km.

Two pyroclastic intervals or complexes primarily composed of tuff-breccias are well-exposed in parts of the central Cagayan Valley and are here designated as marker beds for correlation. These are the Liwan Pyroclastic Complex and the Tabuk Pyroclastics. The Liwan Pyroclastic Complex outcrops along the flanks of Cabalwan and Enrile anticlines. It is generally formed by one or two tuff-breccia deposits from 1 to 3 m thick. The complex is best exposed along the west flank (Figure 9g) of Enrile Anticline just east of the village of Liwan, Kalinga-Apayao Province. In this area, the complex is from 3 to 20 m thick and contains numerous flow units. The Liwan Pyroclastic Complex is distinguished from other pyroclastic deposits by its stratigraphic position along the flanks of anticlines and composition. Tuff breccias of this complex contain rounded lapilli and blocks of pumice in a dacitic matrix in which green-brown hornblende is the principal nonopaque heavy mineral phenocryst. The Tabuk Pyroclastics, flat lying tuff-breccias which cap the Tabuk plateau along the western part of the study area (Figure 9h), are younger and compositionally different than the Liwan Pyroclastic Complex and other tuff-breccias of the Awidon Mesa Formation. Caagusan (1980) recently referred to the Tabuk Pyroclastics as a formation. These deposits are from 8 to 25 m thick and are composed of numerous flow units. Compositionally, the Tabuk Pyroclastics are characterized by subangular lapilli and blocks of dacite and andesite in a dacite matrix which contains oxyhornblende as the principal nonopaque heavy mineral phenocryst.

The pyroclastic deposits of the Awidon Mesa Formation are interbedded in the Cagayan Valley with thicker deposits of polymictic conglomerates, sandstones, siltstones, mudstones, and claystones. Fining upward sequences occur throughout the Formation but are most common in the lower part. Granule to pebble conglomerates and large scale trough cross bedded sandstones fine upward to siltstones and claystones. The sandstones and conglomerates, up to several meters thick, are thinner and more discontinuous than the sandstones and conglomerates of the underlying Ilagan Formation. Lenticular sandstones commonly pinch out within a few hundred meters. Like the sandstones, the associated claystones of the Awidon Mesa Formation are not generally as thick as those of the Ilagan. The upper part of the Awidon Mesa Formation is characterized by thicker, massive polymictic cobble-boulder conglomerates, conglomeritic sandstones, and sandstones. Individual conglomerates are up to 22 m thick while conglomeritic sequences attain a thickness of 50 m. The conglomerates vary from matrix supported to clast supported and are often imbricated indicating flow from the west. The disc to equant shaped rounded clasts are composed primarily of basalt, porphyritic andesite, metasedimentary clasts, quartzite, chert, and a small amount of jasper. Siltstones and claystones occur in this upper conglomeritic sequence but are usually not more than several meters thick and are often truncated by lenticular conglomerates and sandstones.

The conglomerates of the Awidon Mesa Formation in the central Cagayan Valley increase in thickness and clast size from the north to the south.

In the north along the flanks of Cabalwan and Enrile anticline, the conglomerates are up to 6 meters thick and composed of pebble to cobble size clasts. Individual conglomerates form resistant ridges along the flanks of the anticlines and are usually traceable for at least 1 km. Extensive limonite coated lag gravels from erosion of these conglomerates cover many of the hills along the flanks of the anticlines. To the south, at the Wanawan Ranch along the northeastern flank of Pangul Anticline, it was not possible to measure the thickness of the conglomerates because of cover, but an increase in grain size was observed along with an abundance of equant, rounded, small boulders. Farther to the south, the thickest conglomerates occur at the southern nose of Enrile and Pangul anticlines. These conglomerates are up to 50 m thick and coarsen upward from cobbles and small boulders to very large boulders of dacite. The boulders also increase in size to the west from large boulders at southern Enrile Anticline to very large boulders at the south nose of Pangul Anticline, the largest of which has a long axis of 5.5 m.

The lag gravels in the northern part of the study area contain tektites, naturally occurring glasses of possible extraterrestrial or impact origin (Barnes and Barnes, 1973; O'Keefe, 1963). Ninety-one tektites were found during the 1979 field season while thousands had been previously recovered by National Museum field workers. The tektites were once abundant (over 500 were collected at one locality on the Wanawan Ranch by the Philippine National Museum; Lito Soriano, Tuguegarao, Cagayan, personal communication, 1979) but are now scarce as a result of collections made by local people and visitors. Most of the tektites are pebble size, equant in shape, and rounded indicating abrasion by stream transport.

The stratigraphic occurrence of all tektites recovered during this study and previous investigations was noted along with the location of all tektite localities (Figure 12). At this time, it appears that all of the tektites have eroded out of a folded conglomerate or conglomeritic interval stratigraphically beneath the Liwan Pyroclastic Complex along the flanks of Cabalwan, Enrile, and Pangul anticlines. The tektites have not yet been found in situ. The abundance of tektites in some lag gravels suggests that tektites may be found in situ if the underlying conglomerates are excavated and screened.

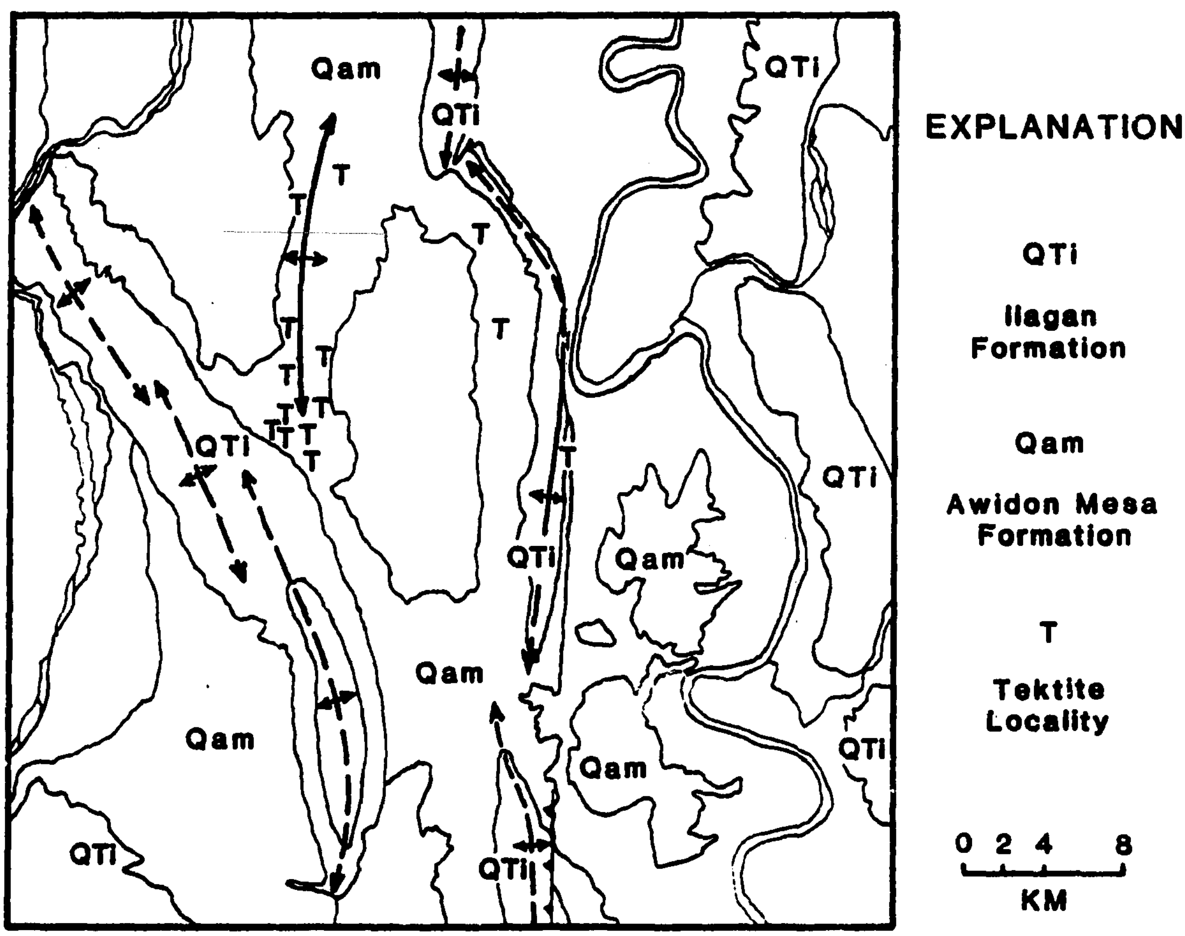


Figure 12. Tektite locality map

The limited stratigraphic occurrence of the tektites throughout the central Cagayan Valley suggests that they are all the same age and may be used for correlation. One tektite from the west flank of Cabalwan Anticline was K-Ar dated as middle Pleistocene, 0.92 $\pm$ .17 Myr. This date roughly correlates with other tektite dates from the Australasian strewn tektite field which average 0.71 Myr (von Koenigswald, 1967). Harrison (1975) has noted that tektite dates must be used with caution because tektites may be easily reworked and deposited with younger sediments. Since the thousands of central Cagayan Valley tektites that have been recovered are from the same stratigraphic position throughout a large area, it is interpreted that they were deposited during a very limited time interval following formation and are not reworked. Radiometric dating of the associated pyroclastic deposits will provide a cross check on the accuracy of the tektite date.

The Awidon Mesa Formation is very fossiliferous in contrast to the underlying Plio-Pleistocene Ilagan Formation. Disarticulated remains of a variety of fossil vertebrates, namely elephants, rhinocerous, carabao, pig, deer, and crocodile occur on the surface and *in situ* in sandstones, conglomerates, and claystones. The fossils occur along the northwest and northeast flanks of Enrile Anticline, the northeast flank of Pangul Anticline, and along both flanks of Cabalwan Anticline. The vertebrates were dated by faunal association as middle Pleistocene by Beyer (1956) and von Koenigswald (1956). Besides vertebrates, permineralized wood and leaf impressions are common. Trace fossils, such as burrows and root casts, are present in many of the siltstones, claystones, and sandstones, and plant molds are common in the pyroclastic deposits.

The middle Pleistocene date assigned to the Awidon Mesa Formation by Durkee and Pederson (1961) was based, in part, on the vertebrate fauna. The tektite date provides additional evidence that the formation is of middle Pleistocene age. Radiometric dates of crystal concentrates from a number of critically located samples were attempted by Professor Frank Fitch of Birkbeck College, University of London. The dates are excessively old (3.7-13.0 Myr) and have a high degree of error ($\pm$ 1.9 to $\pm$ 4.1) because of the low potassium content of the rocks. The minute amounts of radiogenic argon that had been produced in the rocks were, therefore, swamped in dating experiments by argon derived from the atmosphere. This resulted in the excessively old dates, the high degree of error, and even some apparent zero ages (Fitch, University of London, personal communication, 1981). Further attempts to date whole rock samples are planned to try to obtain accurate dates. The whole rock technique was recently used successfully by DeBoer et al. (1980) on rocks of similar composition from central Luzon.

PETROLOGY

The Plio-Pleistocene volcaniclastic sediments of the central part of the Cagayan Valley consist of conglomerates, tuff-breccias and tuffs, sandstones, siltstones, and mudrocks. The tuff-breccias only occur in the Pleistocene Awidon Mesa Formation which also contains a greater proportion of conglomerates than the underlying finer grained Plio-Pleistocene Ilagan Formation.

The sediments display distinct variations in mineralogy, color, cementing agents, and texture. These variations reflect different source rocks and source areas, depositional environments, and post-depositional (diagenetic) alteration. Volcaniclastic sediments are particularly susceptable to rapid alteration by diagenetic processes (Hay, 1957; Pettijohn, 1975; Pittman, 1979; Scholle and Schluger, 1979). This alteration, which includes the dissolution of framework grains and the formation of authigenic clays and zeolites, has considerably changed the original texture and mineralogy of most Plio-Pleistocene volcaniclastic sediments of the Cagayan Valley. Therefore, the composition and diagenetic alteration of the rocks were studied first. This was followed by a textural analysis of samples which had not been significantly altered.

The composition of the sediments was determined by a combination of hand lens, petrographic, X-ray diffraction, and scanning electron microscope methods. Lag gravel and conglomerate clast lithologies were determined by gravel counts of 300 clast fragments identified with a binocular microscope. One hundred fifteen thin sections of conglomerate matrix, sandstones, tuff-breccia matrix, and tuffs were examined with a petrographic microscope. The thin sections were vacuum impregnated with blue epoxy and oil ground when necessary to preserve the original texture and clay minerals. Grain mounts were made of numerous samples which were poorly indurated. A total of 100 points were counted to determine the abundance of major components of each rock, framework grains, detrital matrix, cement types, and porosity. Point counts of 300 framework grains were then made after staining one-half of the slide for feldspars (Bailey and Stevens, 1960). The abundance of framework grain dissolution features was qualitatively estimated using the terms abundant (>10% of the framework grains affected), common (3-10%), sparse (1-3%), and rare (<1%). X-ray diffraction was used to identify sandstone matrix clay, authigenic clay, authigenic zeolite minerals which had been concentrated with heavy liquids, accessory minerals, and the dominant clays of 21 mudrocks. The clays were identified using oriented tile mounts. Fifty-seven and three-tenths mm Debye-Scherrer powder cameras and gelatin mounts (Sorem, 1960) were used to identify accessory minerals. The morphology and occurrence of authigenic minerals and glass shards from selected samples were examined with the scanning electron microscope. The morphology of authigenic clays (Wilson and Pittman, 1977; Neasham, 1977; Stalder, 1973) and zeolites (Mumpton and Ormsby, 1976) is distinctive and useful in mineral identification and interpretation. Each sample was coated with carbon and 400 Å of gold-paladium with a vacuum evaporator to permit electrical conductance and eliminate charging of nonconductive surfaces. Refractive index measurements of glass shards and tektite glass were made using the central focal masking technique (Wilcox, 1979).

Textural characteristics of the samples were determined by a combination of mechanical and sedimentation analyses and microscopic examination. Dominant size, sorting, shape, roundness, and grain contact relationships were noted in thin sections. Twenty-eight conglomerate matrix, tuff-breccia matrix, tuff, and sandstone samples that had not been significantly modified by diagenetic processes were disaggregated with water and sieved into half phi fractions using sieves ranging from -5Ø to 4Ø. The tuff-breccia and tuff samples were sieved by hand for 5 minutes as suggested by Walker (1971) to limit abrasion of pumice grains. The grain size distribution of 10 mudrock samples was determined using the pipette technique. Graphic plots (Glaister and Nelson, 1974) and textural parameters (Folk and Ward, 1957) were then determined so that classification, environmental interpretations, and comparisons between samples could be made.

Conglomerates

All conglomerates in the Upper Member of the Ilagan Formation and the Awidon Mesa Formation are polymictic conglomerates which vary significantly in texture and abundance throughout the study area.

Texture

Granule to pebble conglomerates compose approximately 10% of the Upper Member of the Ilagan Formation. Conglomerates are coarser and more abundant in the Awidon Mesa Formation. Pebble to cobble conglomerates form approximately 20% of the Formation in the northern part of the study area and coarsen to the south where pebble to boulder conglomerates constitute up to 60% of the Formation.

Shape and roundness varies with clast size. The granules are predominantly equant and subangular to subrounded. The pebbles, cobbles, and boulders, in contrast, are equant to disk shaped and rounded to well rounded.

The conglomerate matrix varies from sandy gravel to gravelly sand and muddy sand (Figure 13). The abundance of matrix varies between conglomerates. Most are clast supported, but matrix supported conglomerates with a greater percentage of matrix occur in the Awidon Mesa Formation. Statistical parameters of selected conglomerate samples are recorded in Table 2. The matrix has an average mean diameter of 0.86Ø, coarse sand, and ranges from fine sand (2.86Ø) to granule size (-1.43Ø). The average standard deviation of the samples, 2.19Ø, indicates that the matrix, which varies

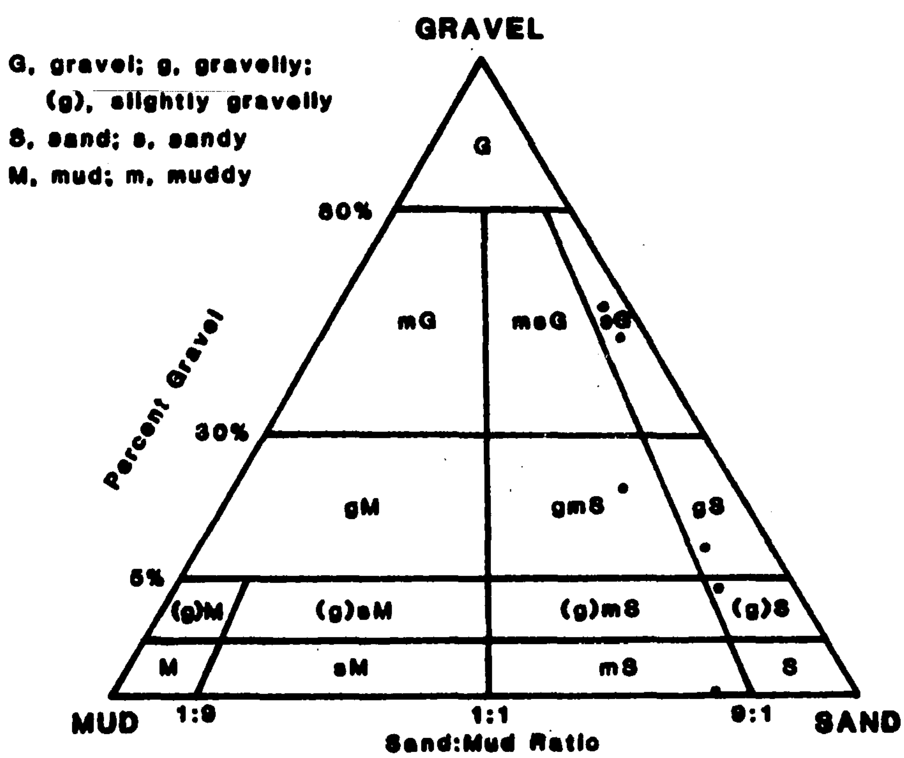


Figure 13. Textural classification of conglomerate matrix samples (after Folk, 1954)

Table 2. Grain size statistics (Folk, 1974) of selected conglomerates, tuff-breccias, tuffs, and sandstones

Sample no.	Phi mean M_Z [a]	Phi deviation σ_I [b]	Phi skewness Sk_I [c]	Phi kurtosis K_G [d]
Conglomerate Matrix				
C4-01	-1.43	3.24	-0.02	0.8
E5A-09b	1.74	1.65	-0.09	1.17
E8-16	2.86	1.15	0.26	1.64
P1-111	1.81	1.84	-0.02	1.98
P4-25	0.81	2.69	-0.19	1.19
M2-02	-0.62	2.56	-0.14	0.89
Tuff-breccia Matrix				
C4A-02a	-1.4	3.14	-0.05	0.69
E1A-15	-0.27	3.10	-0.24	1.07
E2-06a	3.45	2.10	-0.01	1.26
SM1-02c	-2.42	3.24	0.21	0.63
ER1-35b	-0.36	2.59	0.23	0.86
ER1-37a	-0.29	2.45	0.36	0.92
Tuffs				
E1-24	3.71	1.13	0.44	1.02
E6-08	1.81	1.53	0.09	1.15
ER1-21c	4.03	0.85	0.23	0.88
MR1-01b	2.40	1.40	-0.24	0.81
P4-20	3.21	1.30	0.01	1.00
T3-08	4.58	1.55	0.10	0.90
Sandstones				
C4-10	2.47	1.45	0.51	0.17
C5-07	1.41	2.08	0.06	1.39
E2-01	1.80	1.51	0.03	1.18
E6-02	1.62	0.75	0.08	0.99

$$^{a}M_Z = \frac{\phi16+\phi50+\phi84}{3}$$

$$^{b}\sigma_I = \frac{\phi84-\phi16}{4} + \frac{\phi95-\phi5}{6.6}$$

$$^{c}Sk_I = \frac{\phi16+\phi84-2(\phi50)}{2(\phi84-\phi16)} + \frac{\phi5+\phi95-2(\phi50)}{2(\phi95-\phi5)}$$

$$^{d}K_G = \frac{\phi95-\phi5}{2.44(\phi75-\phi25)}$$

Table 2 (continued)

Sample no.	Phi mean M_Z[a]	Phi deviation σ_I[b]	Phi skewness Sk_I[c]	Phi kurtosis K_G[d]
ER1-33	1.73	0.69	0.22	1.13
ER3-09	1.61	1.64	0.21	1.86
M2-03	1.47	1.55	0.28	1.31
P4-31	3.39	2.37	0.41	1.21
T3-02a	1.35	1.40	0.09	1.49
T3-07a	1.67	1.49	-0.26	1.21

from 1.15Ø (poorly sorted) to 3.24Ø (very poorly sorted), is typically very poorly sorted. Skewness of the matrix particle size distribution, -0.03, is nearly symmetrical with a range from fine skewed (0.26) to coarse skewed (-0.19). The average kurtosis value of the matrix, 1.28, indicates that it is leptokurtic, i.e. better sorted in the central portion than in the extremes. Kurtosis ranges from platykurtic (0.80) to very leptokurtic (1.98).

The coarse grain size, very poor sorting and variations in clast size, reflect the textural immaturity of the conglomerates and variations in depositional processes. The more abundant and coarser conglomerates of the Awidon Mesa Formation were deposited by higher energy streams than those of the Ilagan. The coarsening in the Awidon Mesa Formation from pebble to cobble size clasts in the eastern and northern parts of the study area to cobbles and boulders in the south and west reflects the development of high energy braided streams and debris flows in the Pleistocene. Probability plots of Awidon Mesa Formation conglomerate matrix samples (Figure 14) display textural variations that are typical of braided stream and debris

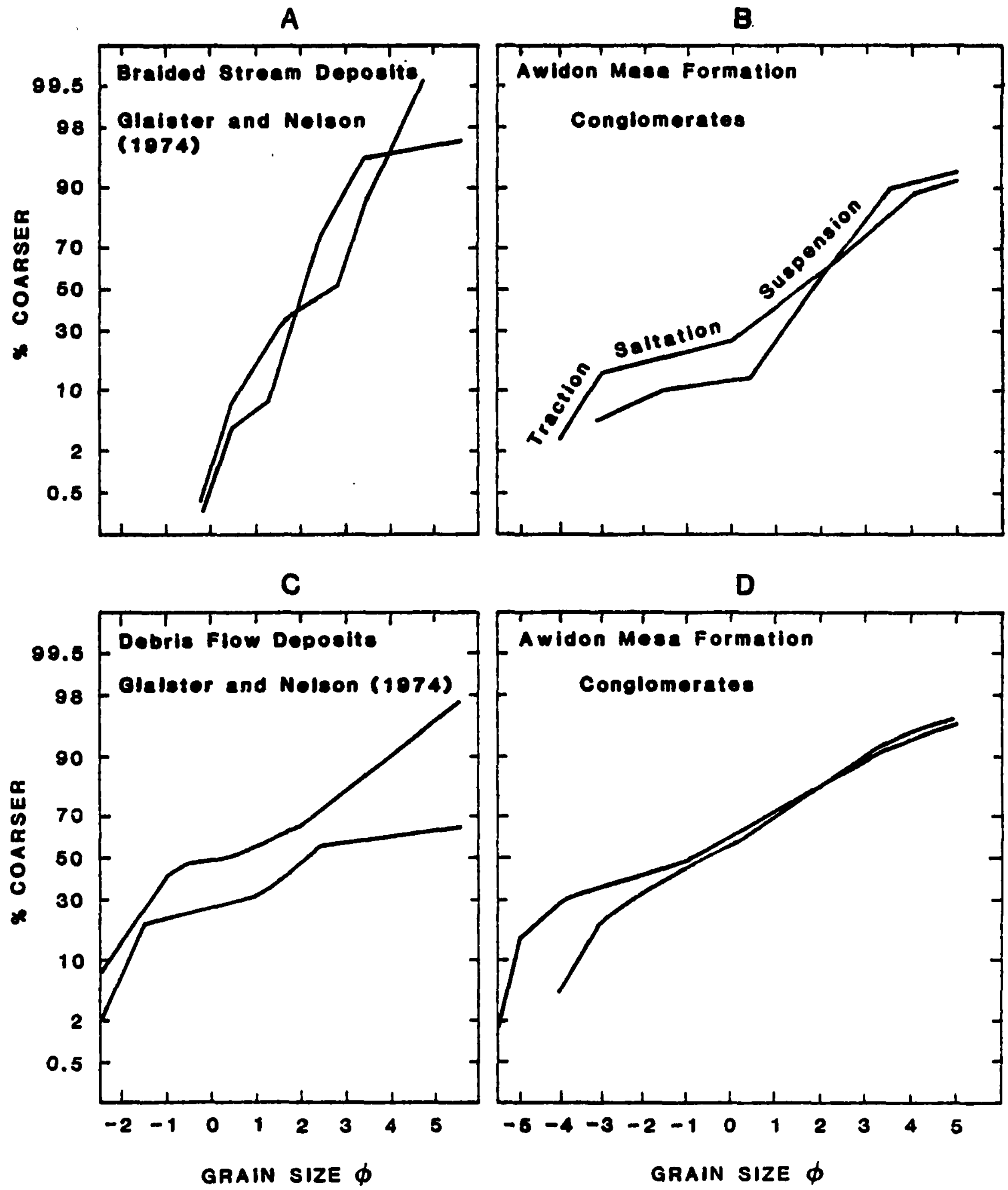


Figure 14. Probability plots of braided stream and debris flow deposits (Glaister and Nelson, 1974) and plots of Awidon Mesa Formation conglomerates interpreted as braided stream and debris flow deposits

flow deposits as described by Glaister and Nelson (1974). The braided stream deposits are strongly bimodal, have sharp saltation and traction junctions, and have a low saltation population which is expressed by a low slope denoting poor sorting. Debris flow deposits, in contrast, plot as a broken line with a very low slope indicative of extremely poor sorting. These deposits, which constitute up to 60% of the Awidon Mesa Formation in the southern and western parts of the study area, constitute the major part of an alluvial fan which formed at the base of the Cordillera Central as a result of Pleistocene uplift.

Composition

Clast lithologies are similar in both the Ilagan and Awidon Mesa formations. Porphyritic andesite and basalt are the dominant lithologies recorded in pebble counts of selected conglomerates and river gravels (Table 3). Metasedimentary clasts and quartzite are common components while jasper, agglomerate (welded tuff clasts with fiamme), metamorphic clasts, chalcedony, diorite, hydrothermally altered and mineralized rock fragments occur in minor amounts. The basalt, andesite, and diorite clasts are moderately to extensively altered. Diorite was not generally differentiated from the altered basalt in pebble counts because the alteration made distinction difficult. Nearly all the clasts exposed at the surface or in lag gravels are coated with a limonite crust or rind.

In the upper part of the Awidon Mesa Formation, tektites occur in lag gravels which are derived from one pebble conglomerate or a conglomeritic sandstone interval. The tektites, which are the same size as the associated pebbles, are dominantly spherical and rounded (Figure 15). Rounded

Table 3. Clast lithologies of selected conglomerates and river gravels

Sample no.	Andesite	Basalt	Metasediment	Quartzite	Jasper	Chalcedony	Agglomerate	Diorite	Metamorphic	Other
Awidon Mesa Formation										
C1-5	39	34	4	9	-	-	-	-	2	12
C5-32	29	28	24	8	1	1	1	-	-	8
C4-31	26	25	13	11	2	-	5	-	6	3
ER1-17	38	48	4	3	-	-	1	-	3	3
ER1-34	53	26	17	2	1	-	1	-	-	-
MR1-01	51	32	3	8	-	-	1	-	6	-
WR1-12a	49	11	15	13	-	1	-	-	3	8
Ilagan Formation, Upper Member										
P1-31	37	30	20	11	-	-	-	-	2	-
P1-91	57	23	5	9	-	-	-	-	1	3
River Gravels										
Chico R. Tabuk	41	34	15	3	-	-	-	5	1	-
Cagayan R. Tuguegarao	53	28	7	4	1	-	-	3	1	3

teardrop and dumbbell shaped tektites (Figure 15d and e) also occur but are rare. The rounding suggests that the tektites were abraded during stream transport along with the associated pebbles. A variety of surface textures, grooves, wrinkles, conchoidal fractures, pitts and vesicles are present on the tektites as illustrated in Figure 15. Flow structures in the tektite glass are also observable on the surface and in thin sections (Figure 15h and i). The grooved and pitted surface sculpture and flow structures are typical of Philippine (Beyer, 1961) and Southeast Asia tektites (O'Keefe, 1963). The composition of a tektite from the west flank of Cabalwan

Figure 15. Morphology of central Cagayan Valley tektites.

a. Spherical rounded tektite which was rounded during stream transport

b. Spherical tektite with wrinkled surface and deep grooves

c. Spherical tektite with conchoidal fractures on surface

d. Dumbbell shaped tektite with flow structures visible on surface

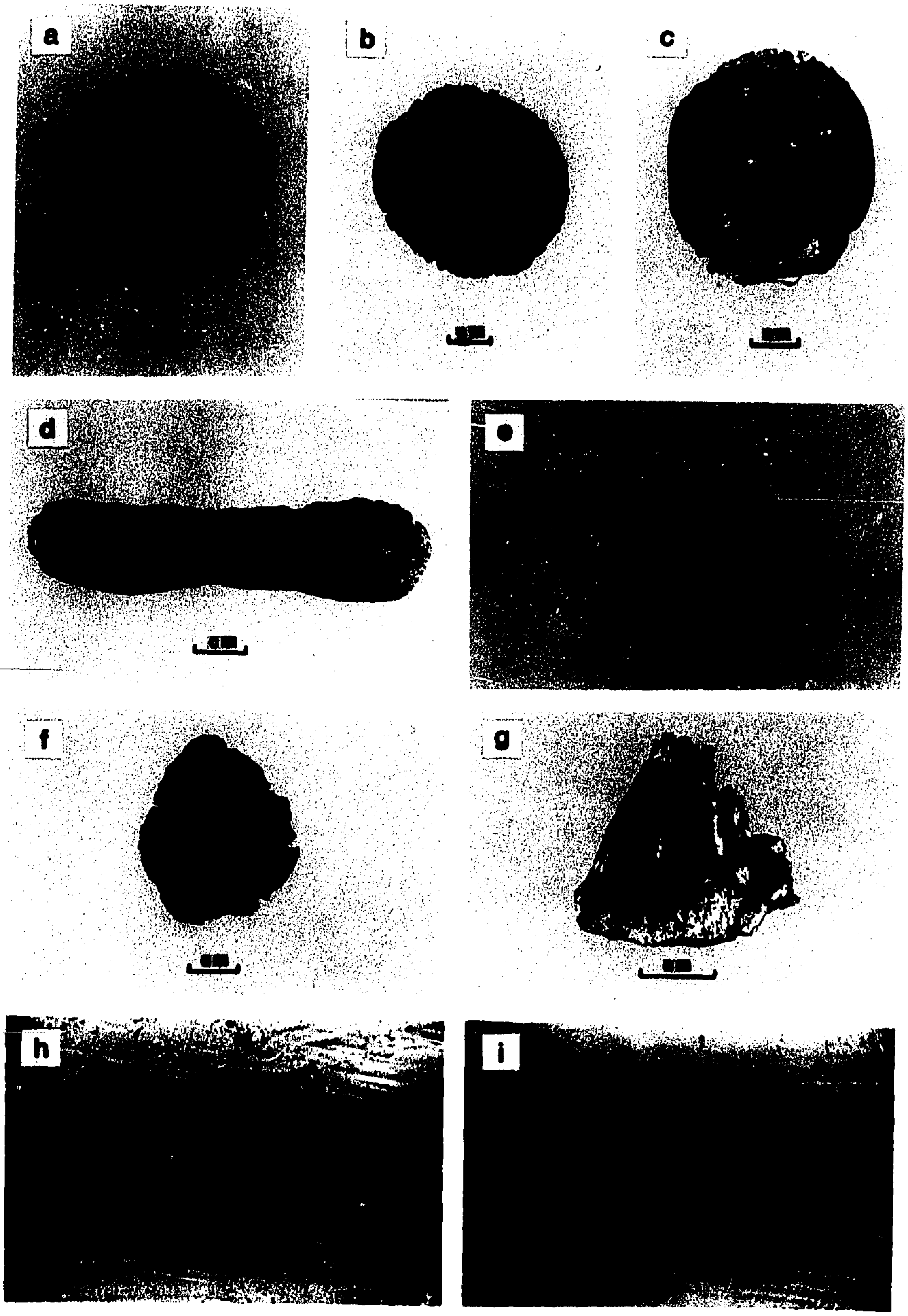
e. Teardrop shaped tektite with pitted vesicular surface

f. Vesicular tektite fragment with deep grooves in outer surface

g. Tektite fragment exhibiting radial fractures from center of tektite to surface. The fractures may be contraction cracks formed during rapid cooling of the tektite

h. Photomicrograph of tektite flow structure

i. Photomicrograph of flow structure deflected around vesicle


a
b
c
d
e
f
g
h
i

Anticline was analyzed by Vondra et al. (1981) and correlates with the average Philippine tektite composition (Table 4). The Cabalwan Anticline tektite glass is greenish, undevitrified, and has a refractive index of 1.511. This value also correlates with other Philippine tektite refractive index measurements which range from 1.5081 to 1.5191 (Chao, 1963, p. 59).

Table 4. Major and minor element composition of a Cagayan Valley tektite (Cabalwan Anticline) and average Philippine tektite composition (From Vondra et al. (1981). Cagayan Valley tektite analyzed by D. Burggraf)

	Tektite composition[a] Cagayan Valley	Average Philippine tektite composition[a]	
		Barnes & Barnes (1973, p. 106)	O'Keefe (1963, p. 69)
SiO_2	68.82	71.21	70.80
Al_2O_3	14.31	12.57	13.85
FeO_T	5.30	5.51	4.93
CaO	2.49	3.19	2.89
MgO	2.27	2.90	2.75
Na_2O	1.62	1.52	1.78
K_2O	2.50	1.93	2.35
MnO	0.10	0.11	0.09
TiO_2	0.84	0.89	0.75

[a]Oxide weight percent.

Compositionally, the conglomerate matrix ranges from litharenite to arkose with most samples classified as feldspathic litharenites (Figure 16). The matrix composition, percent framework grains, detrital matrix, cement and porosity, and mineralogical components of the framework grains as determined from thin sections are recorded in Table 5 along with

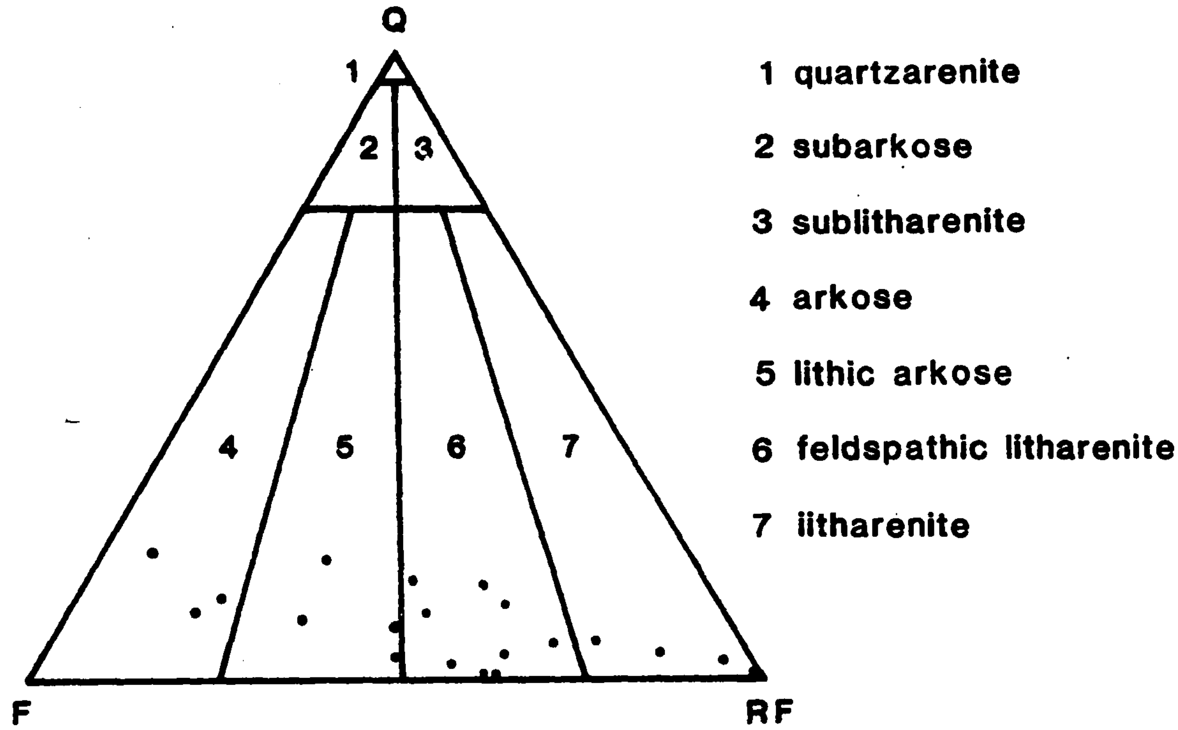


Figure 16. Compositional classification of conglomerate matrix samples (after Folk, 1974)

qualitative estimates of the extent of grain dissolution. Framework grains, which constitute 47 to 66% of the matrix, are dominantly composed of sodic plagioclase, andesine, and volcanic rock fragments. Quartz, sedimentary rock fragments, hornblende, magnetite, and ilmenite are constant minor minerals. The quartz characteristically occurs as coarse sand to granule size bipyramidal crystals. Potassium feldspar, metamorphic and plutonic igneous rock fragments, oxyhornblende, augite, and hypersthene also occur but are generally rare. Detrital matrix (Protomatrix of Dickinson, 1970a) varies in abundance from less than 1% to 47%. Authigenic cements also occur in most conglomerates. Clay rims and clay coats composed of smectite commonly coat the framework grains and constitute as much as 17% of the samples. Authigenic zeolites, primarily clinoptilolite

Table 5. Conglomerate matrix composition and mineralogical components of the framework grains

	Matrix composition						
			Cement				
Sample no.	Framework grains	Detrital matrix	Clay	Calcite	Zeolites	Dissolution features[a]	Porosity
Awidon Mesa Formation							
C1-05	GM[b]						
C2-01	55	2	17	-	-	A	26
C4-01	47	33	-	-	-	C	20
C4-39	GM						
E3-03	GM						
E5-01a	GM						
E5-09b	47	47	-	-	-	R	6
E6-04a	GM						
E7-01	GM						
E8-01a	58	1	7	-	22	C	12
E8-16	56	21	1	-	-	R	22
ER1-17	GM						
ER1-32	60	-	13	-	-	C	27
P1-111	52	21	-	-	-	A	27
P4-25	66	6	4	-	-	S	24
M2-02	55	30	2	-	-	R	13
M3-01a	GM						
SM1-07b	61	1	11	-	-	C	27
Ilagan Formation							
P1-91	56	1	12	-	-	A	31
P1-16	57	2	4	-	27	A	10

[a]A=abundant; C=common; S-sparse; R=rare.

[b]Grain mount; not possible to point count original composition.

[c]t represents less than 0.5%.

Framework grain mineralogy										
			Rock fragments			Amphiboles		Pyroxenes		
Quartz	K. feldspar	Plagioclase	Volcanic	Sedimentary	Metamorphic, plotunic igneous	Hornblende	Oxyhornblende	Augite	Hypersthene	Opaque minerals
6	-	23	49	17	t[c]	1	-	-	-	4
3	-	40	48	3	3	3	-	-	-	-
8	-	29	28	1	6	-	-	7	4	17
9	-	50	30	-	-	8	-	-	-	3
7	-	39	38	-	-	7	-	-	-	9
-	-	32	50	2	-	-	-	7	5	4
16	-	52	4	-	-	11	1	1	-	15
6	-	25	63	3	-	1	-	-	-	2
-	-	-	64	36	-	-	-	-	-	-
14	-	27	46	1	1	2	-	6	1	2
17	-	42	25	-	1	9	-	-	-	6
4	-	31	55	-	1	1	-	6	-	2
-	-	29	47	2	1	1	-	4	3	13
4	1	43	11	12	20	3	-	t	-	6
12	-	27	51	2	1	1	-	-	-	6
15	-	37	37	-	4	2	-	-	-	5
11	-	64	12	1	1	6	-	-	-	5
12	-	58	15	1	-	3	-	-	-	11
8	-	12	75	1	2	1	-	-	-	1
4	1	3	90	-	-	-	-	2	-	-

and stilbite are abundant in two samples where they compose up to 27% of the rock. The average intergranular and dissolution porosity of the conglomerate matrix is 20%. Dissolution of volcanic rock fragments, plagioclase, amphiboles, and pyroxenes has occurred in all samples affecting from approximately 1 to 50% of the grains. The mineralogical composition, cements, and dissolution features of the conglomerate matrix are similar to the sandstones and are described in more detail in the discussion of sandstones.

The composition of the conglomerates indicates that they are derived from the adjacent volcanic arcs. The basalt, porphyritic andesite and metasedimentary clasts may be derived from the Cordillera Central and/or the Sierra Madre volcanic arc. Bedded red cherts occur in both the Cordillera Central and the Sierra Madre (Durkee and Pederson, 1961; Caagusan, 1980) and are the probable sources of the jasper. The unstable primary volcanic minerals in the matrix also reflect the composition of the adjacent arcs. The greater percentage of hornblende and bipyramidal quartz suggests, however, that the Cordillera Central was the dominant source.

Pyroclastic Rocks

Texture

The pyroclastic rocks of the central part of the Cagayan Valley may be classified as tuff-breccias, breccias with an abundant matrix of ash-size fragments (Fisher, 1966; Willams and McBirney, 1979), and tuffs. Volumetrically, tuff-breccias are more abundant than tuffs. The tuff-breccias, however, are limited to the Awidon Mesa Formation while tuffs occur in both the Ilagan and Awidon Mesa formations.

The tuff-breccias vary in block size and roundness according to their geographic occurrence and age. Tuff-breccias that occur in the anticlines contain rounded lapilli, fine blocks, and occasional coarse blocks of pumice. The younger tuff-breccias that underlie the Tabuk plateau contain subangular lapilli to coarse blocks of andesite and dacite. Some deposits, however, are characterized by subrounded to rounded blocks. Rounded blocks of dacite as large as 1.5 m (long axis) occur at the south nose of Enrile Anticline while blocks up to 5.5 m occur at the south nose of Pangul Anticline.

Matrix is the dominant component of the tuff breccias, usually composing more than 50% of the rock. Statistical parameters of the size-frequency distribution of selected tuff-breccias are recorded in Table 2. The average mean diameter is very coarse ash (-0.22Ø) and varies from very fine ash (3.45Ø) to fine lapilli (-2.42Ø). The matrix is typically very poorly sorted with an average standard deviation of 2.77Ø. Skewness varies from strongly fine skewed (0.36) to coarse skewed (-0.24) and is characterized by an average value of 0.04, nearly symmetrical. The average kurtosis value is 0.90, mesokurtic, but values range from 0.63, very platykurtic, to 1.26, leptokurtic.

Probability plots of tuff-breccia size analyses are similar to plots of density flow and debris flow deposits (Figure 17) as compiled by Glaister and Nelson (1974). Density flow deposits often plot as a broken convex line which runs at a low to moderate slope indicative of poor sorting (Figure 17a). The arcuate shape of the curves probably reflects sorting conditions within the suspension (Glaister and Nelson, 1974) as heavier grains respond to gravitational segregation more efficiently than

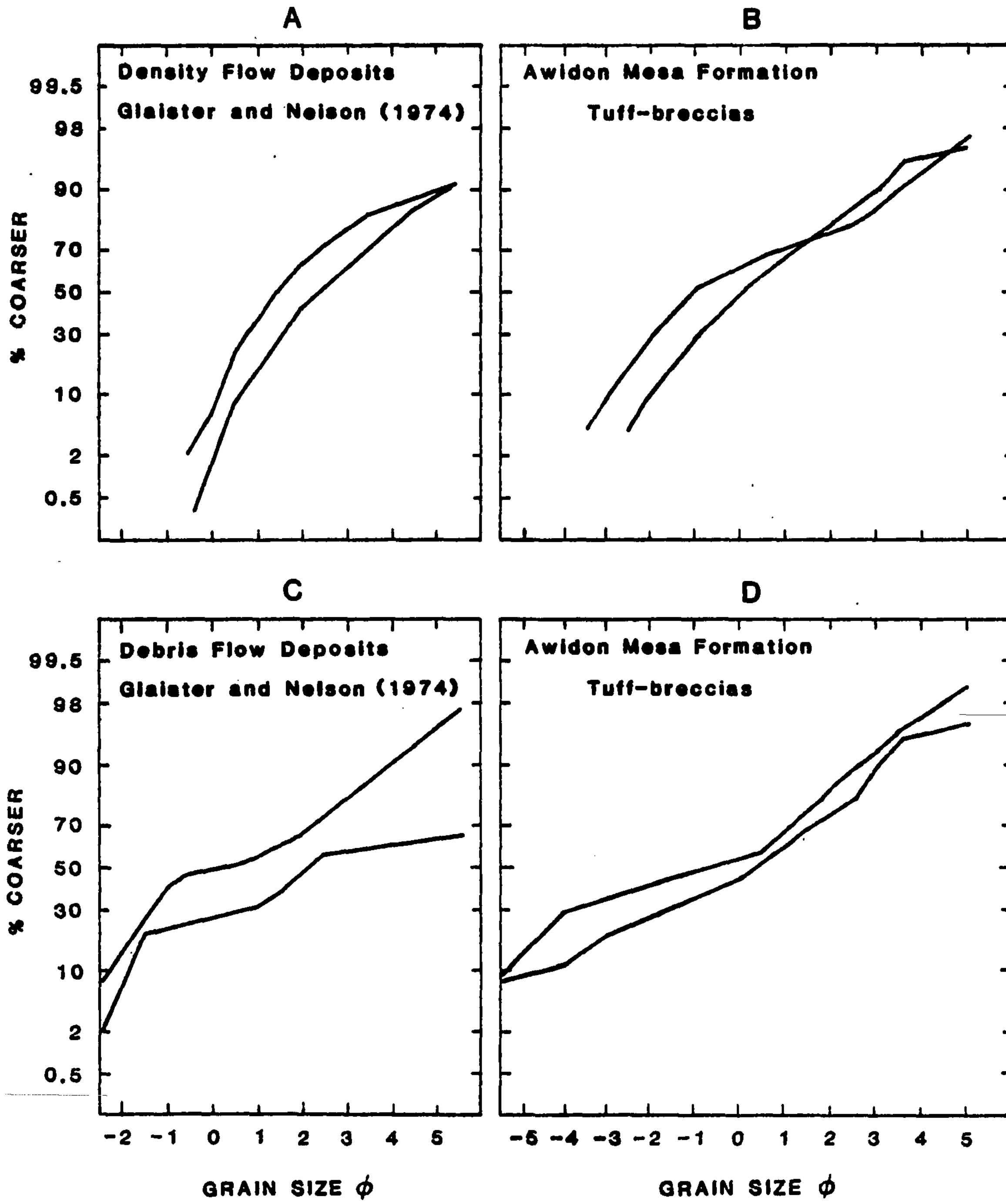


Figure 17. Probability plots of density and debris flow deposits (Glaister and Nelson, 1974) and Awidon Mesa Formation tuff-breccia samples interpreted as density and debris flow deposits

lighter grains which are kept in suspension by turbulence. Plots of debris flow deposits (Figure 17c) have an erratic curve of very low slope indicative of very poor sorting. The curve segments have no genetic significance, and the very poor sorting suggests the lack of any transport mechanism capable of size grading (Glaister and Nelson, 1974). The similarity of tuff-breccia curves (Figure 17b and d) with density flow and debris flow plots (Figure 17a and c) suggests that the tuff-breccias were deposited as density and debris flows.

Textural parameters of selected tuffs are presented in Table 2. The average mean grain size diameter of the analyzed tuffs is 3.29Ø, very fine ash, and ranges from 1.81Ø, very coarse ash, to 4.58Ø, very coarse silt. Numerous finer grained tuffs and tuffaceous mudrocks also occur in the Plio-Pleistocene sediments. The average standard deviation is 1.29Ø, poorly sorted, and varies from 1.55Ø, poorly sorted, to 0.85Ø, moderately sorted. The tuffs range from coarse skewed (-0.24) to strongly fine skewed (0.44) and average 0.11, fine skewed. Kurtosis values are dominantly mesokurtic (0.96) and range from leptokurtic (1.15) to platykurtic (0.81). Probability plots of the tuffs are similar to the density flow deposits of Glaister and Nelson (1974) (Figure 18), which suggests the tuffs were deposited entirely from suspension.

Walker (1971), Sparks (1976), and Blake (1976) have demonstrated that plots of two textural parameters, MdØ and σØ of Inman (1952), may be used to differentiate between rocks deposited by pyroclastic flows or falls. To facilitate comparison with the published grain size data collected at whole phi intervals, textural analyses of the Cagayan Valley pyroclastic rocks were plotted at whole phi intervals, and the Inman (1952) parameters MdØ

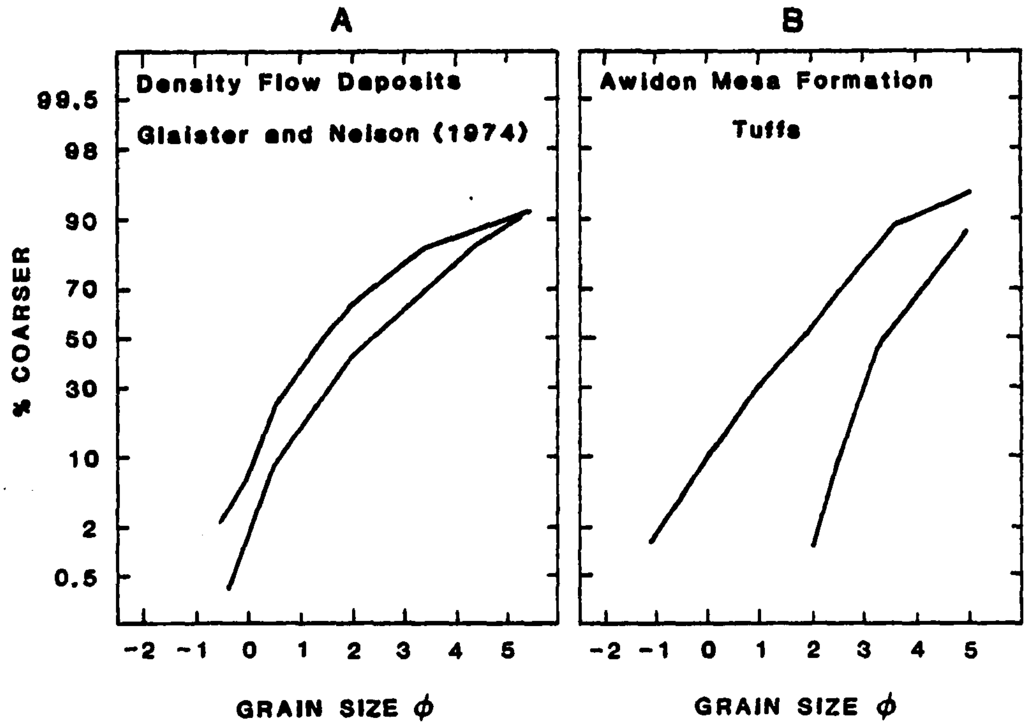


Figure 18. Prc bability plots of Cagayan Valley tuffs and density flow deposits as compiled by Glaister and Nelson (1974)

and σØ were calculated. The tuffs plot in the pyroclast fall field defined by Walker (1971), while all tuff-breccia samples except one plot in the flow field (Figure 19). The tuff-breccia that plots outside the flow field has a slightly larger mean clast size value.

Textural data suggest that the pyroclastic rocks are of both pyroclast fall and flow origin. The tuff-breccias are pyroclast flow deposits which were transported as density and debris flows. The finer grained tuffs which were better sorted than the tuff-breccias are interpreted as fall deposits which were deposited from suspension. Field evidence and compositional data indicate that many of the fall deposits are not primary but have been reworked to varying degrees.

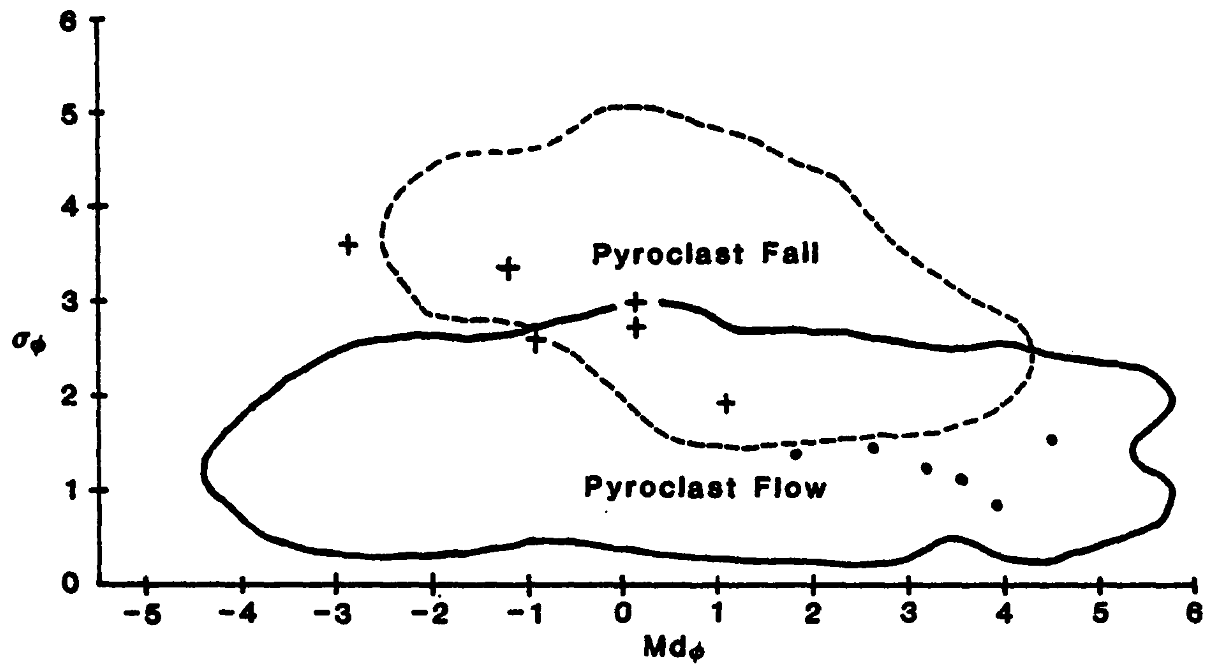


Figure 19. Mdϕ/$\sigma\phi$ plot of Awidon Mesa Formation pyroclast fall and flow deposits. (Contours (1% contours) for pyroclast fall and flow fields are based on 1,600 analyses compiled by Walker (1971))

Composition

Distinct variations occur in the composition of both the tuff-breccias and tuffs. The composition of the tuff-breccia matrix and tuffs is recorded in Table 6.

Tuff-breccia composition varies significantly between the older folded deposits of the anticlinal belt (including the Liwan Pyroclastic Complex) and the younger flat-lying deposits of the Tabuk plateau, the Tabuk Pyroclastics. The folded tuff-breccias are composed of pumice blocks and lapilli in a dacitic tuff matrix which is composed primarily of plagioclase (55%), bipyramidal quartz (9%), volcanic rock fragments (8%), pumice grains (14%), green-brown hornblende (10%), and opaque minerals, magnetite, and ilmenite (4%). The younger Tabuk Pyroclastics, in contrast, are composed

Table 6. Composition of tuff-breccia matrix and tuffs

	Composition					
			Cements			
Sample no.	Framework grains	Matrix	Clay	Zeolites	Dissolution features[a]	Porosity
TUFF-BRECCIAS						
Liwan Pyroclastics						
C1-03b	54	35	-	-	-	11
C2-03b	48	48	-	-	-	4
C4A-02a	37	54	-	-	-	9
C5-33	GM[b]					
E1A-15	45	31	2	-	R	22
E2-03a	53	-	-	-	-	47
E2-06a	GM					
E3-05b	57	23	-	-	-	20
E5-10a	56	37	-	-	S	7
E6-01	46	21	-	-	-	33
E7-10	53	22	2	-	-	23
E8-20	51	34	2	-	-	13
M1-01	62	31	-	-	-	7
Tabuk Pyroclastics						
P2-07b	37	44	-	-	-	19
T2-03	GM					
T3-10	38	53	-	-	-	9
T3-10b	55	36	-	-	-	9

[a]A=abundant; C=common; S=sparse; R=rare.

[b]Grain mount.

[c]t represents less than 0.5%.

		Framework grain mineralogy								
		Rock fragments				Amphiboles		Pyroxenes		
Quartz	Plagioclase	Volcanic	Pumice or glass shards	Sedimentary	Metamorphic, plutonic igneous	Hornblende	Oxyhornblende	Augite	Hypersthene	Opaque minerals
9	56	9	9	3	-	9	-	-	-	5
11	53	8	10	1	1	13	-	-	-	3
14	52	6	13	-	-	13	-	-	-	2
9	52	12	11	-	1	13	-	-	-	2
15	49	11	19	-	-	5	-	-	-	1
10	66	-	-	-	-	21	-	-	-	3
7	58	5	11	-	-	13	-	-	-	6
7	53	2	26	3	-	8	-	-	-	1
8	66	10	6	-	t[c]	7	-	-	-	3
5	38	4	46	-	-	7	-	-	-	t
14	56	3	15	-	-	11	-	-	-	1
11	61	5	13	-	-	10	-	-	-	t
6	56	2	27	-	-	9	-	-	-	t
5	64	-	-	-	-	-	23	1	-	7
6	37	51	-	-	-	t	4	-	-	2
7	49	25	-	-	1	-	12	t	-	6
7	72	-	-	-	-	-	19	-	-	2

Table 6. (continued)

Sample no.	Composition: Framework grains	Matrix	Cements: Clay	Cements: Zeolites	Dissolution features[a]	Porosity
Other Deposits						
E5-30	45	28	1	-	S	26
ER1-35b	65	25	1	-	C	8
ER1-37a	46	37	1	-	S	16
P2-04	50	32	2	-	S	16
P4-08	60	31	-	-	C	9
P4-33	55	33	-	-	S	12
SM1-02c	56	29	-	-	S	15
TUFFS						
Awidon Mesa Formation						
E1-24	60	22	-	-	-	18
E3A-07b	68	-	2	-	-	30
E6-08	59	25	-	-	S	16
E8-05	GM					
ER1-21c	50	26	1	-	-	23
ER4-12	29	57	3	-	-	11
P3-10	60	1	19	-	S	20
P4-20	68	19	3	-	-	10
T3-08	45	25	-	-	-	30
Ilagan Formation						
E1-43	44	15	-	31	C	9
P1-51a	61	-	-	37	S	2
P1-70	60	1	1	22	A	16

Framework grain mineralogy										
		Rock fragments				Amphiboles		Pyroxenes		
Quartz	Plagioclase	Volcanic	Pumice or glass shards	Sedimentary	Metamorphic, plutonic igneous	Hornblende	Oxyhornblende	Augite	Hypersthene	Opaque minerals
2	43	31	9	3	-	8	-	1	-	3
4	48	18	9	1	-	17	-	1	-	2
-	-	-	100	-	-	-	-	-	-	-
8	56	8	8	1	2	9	-	-	-	8
4	58	15	4	-	-	5	-	3	-	11
22	65	-	-	-	-	12	-	-	-	1
7	67	6	10	-	-	8	-	-	-	2
4	64	20	-	-	-	-	4	-	-	8
-	20	-	78	-	-	2	-	-	-	-
8	62	9	5	-	-	12	-	-	-	4
-	43	48	-	2	1	2	-	2	-	2
1	67	9	8	-	-	9	1	-	-	5
3	72	-	-	-	-	16	-	-	-	9
7	47	23	-	-	3	7	-	4	1	8
-	20	-	80	-	-	-	-	-	-	-
5	69	6	-	-	-	-	10	-	-	10
4	66	20	-	7	2	1	-	-	-	-
1	17	80	-	-	-	-	-	2	-	-
3	40	28	-	1	-	3	-	4	t	21

of andesite and dacite blocks and lapilli in a matrix of plagioclase (55%), bipyramidal quartz (6%), oxyhornblende (14%), and opaque minerals (4%) with a variable percentage of volcanic rock fragments (0-51%). The major difference in the composition of the matrix is the pumice and heavy mineral content of the rocks. The Tabuk Pyroclastics lack pumice and hornblende but contain a significant amount of oxyhornblende in contrast to the older tuff-breccias.

The tuff-breccia matrix is composed primarily of framework grains, 51%, with 32% matrix and 16% intergranular porosity. The framework grains are primarily phenocrysts, but microlitic volcanic rock fragments (Dickinson, 1970a) and pumice grains or glass shards are also common in many samples. The phenocrysts are dominantly euhedral (Figure 20a). Plagioclase grains are strongly zoned and commonly display albite and carlsbad twins. Extinction angles of albite twin planes indicate that the composition is in the andesine range. The bipyramidal quartz crystals are water clear and typically have planar crystal faces but are often rounded indicating resorption (Figure 20c). Oxyhornblende grains are almost all euhedral while the hornblende grains are dominantly subhedral. Ilmenite and magnetite often occur as octahedral crystals. Dissolution of some plagioclase and heavy mineral phenocrysts has occurred in numerous tuff-breccias but is very minor. The finer grained matrix varies in composition. Most matrix is formed by silt-size angular fragments, primarily feldspar, and occasional glass shards (Figure 20b), but feldspar laths may also form the matrix. Amorphous silica, with a modal refractive index of 1.518, forms a cement in some samples such as Figure 20c and d which contain a

Figure 20. Photomicrographs of Cagayan Valley pyroclastic rocks

a. Tuff-breccia, P4-33, containing euhedral phenocrysts of plagioclase, bipyramidal quartz, and hornblende

b. Tuff-breccia matrix, E3-05b, formed by angular feldspar fragments and glass shards

c. Tuff-breccia matrix, P2-07b, formed by feldspar laths and amorphous silica cement

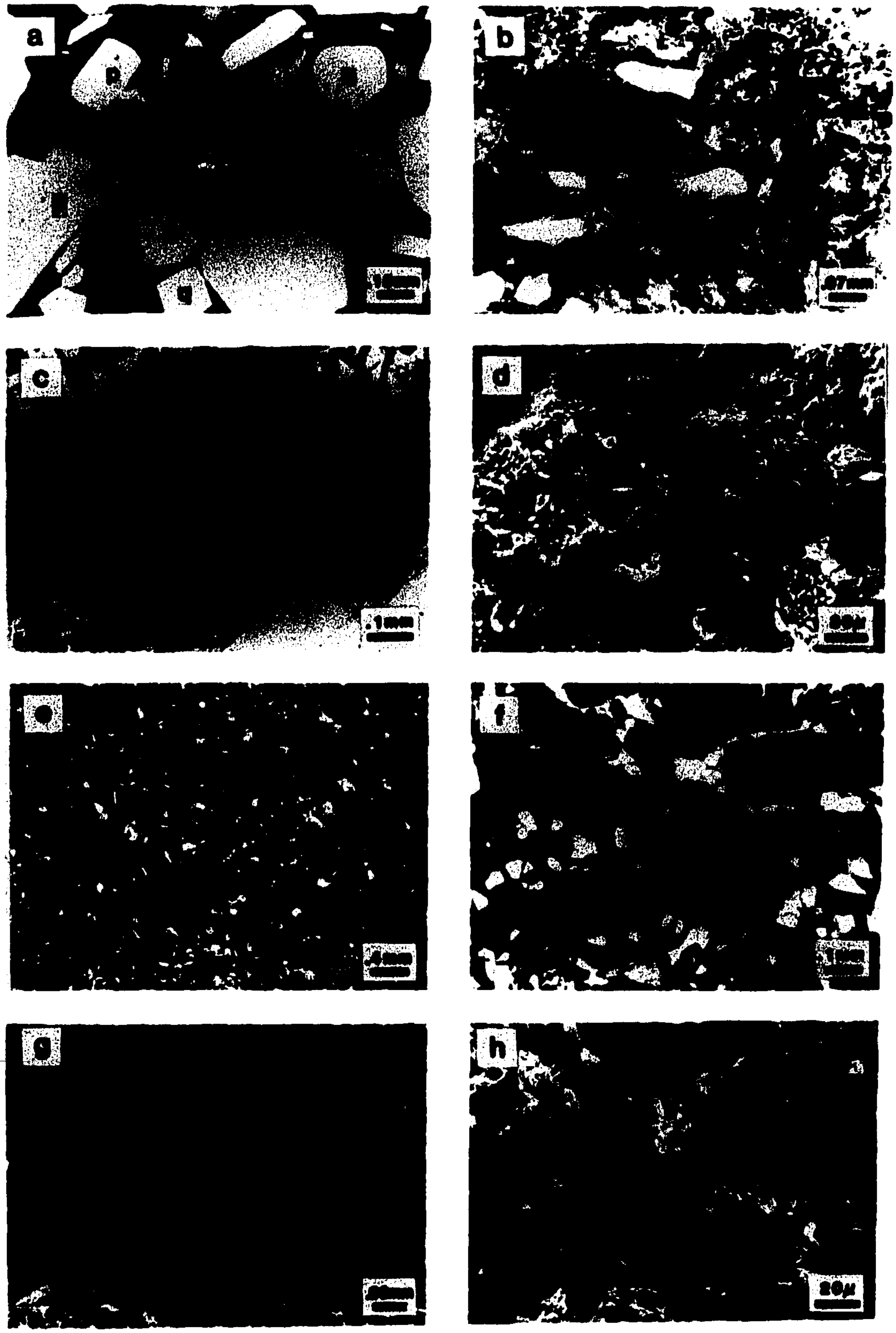
d. SEM photomicrograph of amorphous silica coating tuff-breccia matrix, P2-07b

e. Crystal tuff, ER1-21c, composed primarily of angular crystal fragments

f. Lithic tuff, P1-51a, composed of microlitic volcanic rock fragments and zeolite (stilbite) cement

g. Vitric tuff, E3A-07b, composed of glass shards with Y-shaped junctures and angular feldspar fragments

h. SEM photomicrograph of vitric tuff glass shards and feldspar fragments, E3A-07b


a
b
c
d
e
f
g
h
20μ

matrix of feldspar laths. Authigenic clay, primarily smectite, is a very minor cement in some tuff-breccias.

The tuffs are predominantly crystal tuffs, but vitric and lithic tuffs also occur (Figure 21). Angular plagioclase grains are the dominant component of the crystal tuffs (Figure 20e) while volcanic rock fragments, bipyramidal quartz, glass shards, hornblende, oxyhornblende, augite, hypersthene, magnetite, and ilmenite are common minor constituents. The vitric tuffs consist of glass shards with a typical Y-shape characteristic of bubble wall junctures (Figure 20g and h) and angular plagioclase crystals. The glass shards are dominantly fresh to slightly altered. The lithic tuffs (Figure 20f) are composed of microlitic volcanic rock fragments (Dickinson, 1970a) and lesser amounts of crystals or glass shards and pumice

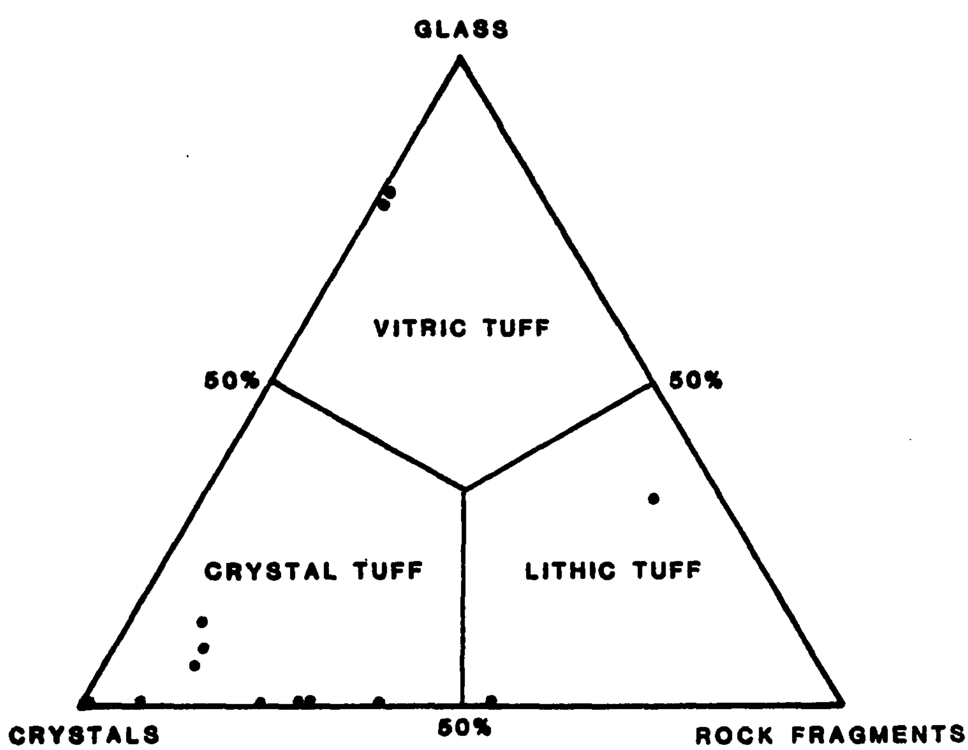


Figure 21. Classification of Cagayan Valley tuffs (after Pettijohn, 1957)

grains. Most of the tuffs contain a variable amount (0-57%) of finer grained matrix and average 17% porosity.

Pumice grains and glass shards occur only in tuffs and tuff-breccias of the Awidon Mesa Formation. All are colorless and have similar modal refractive indices (Table 7). All samples from the Liwan Pyroclastic Complex have indices between 1.499 and 1.503. Samples from other deposits vary between 1.499 to 1.502.

Table 7. Modal refractive indices of Awidon Mesa Formation tuff and tuff-breccia pumice fragments or glass shards

Sample no.	Modal refractive index
Liwan Pyroclastic Complex	
C1-03b	1.501
C4A-02a	1.502
C5-33	1.499
E1A-15	1.500
E3A-07b	1.503
E7-10	1.501
E8-20	1.500
Other Pyroclastic Deposits	
ER1-35b	1.499
ER1-37a	1.502
P4-33	1.500
SM1-02c	1.500

The amount of alteration of the tuffs varies significantly between the Upper Member of the Ilagan Formation and the Awidon Mesa Formation. The Ilagan Formation tuffs contain approximately 30% zeolite cement, clinoptilolite, and/or stilbite, which completely cements some tuffs (Figure 20f), reducing porosity to as low as 2%. Dissolution of plagioclase, heavy minerals, and volcanic rock fragments is also pronounced in the Upper

Member tuffs which lack glass shards. The Awidon Mesa tuffs, in contrast, contain significant amounts of fresh to slightly altered glass and are not significantly altered by dissolution of framework grains or zeolite cements. Significant amounts of authigenic smectite do occur, however, in some samples. The absence of glass shards and the presence of zeolite cements in Upper Member of the Ilagan Formation suggests that the zeolites may have formed by alteration of the original volcanic glass which was probably present in the Ilagan Formation. The alteration of volcanic glass to zeolite minerals is a common diagenetic reaction in volcaniclastic rocks (Hay, 1966; Surdam and Boles, 1979).

Preliminary chemical analyses of the pyroclastic deposits (D. Burggraf, personal communication, 1981) indicate that the pyroclastic rocks are intermediate to felsic in composition (53.96 to 70.01% SiO_2). There is a distinct increase in silica content from the Plio-Pleistocene Ilagan Formation (56.54%) to the Pleistocene Awidon Mesa Formation (65.18%). The K_2O content of the pyroclastic rocks also increases from the Ilagan Formation (0.83%) to the Awidon Mesa Formation (1.23%). The increase in silica and potassium content of the pyroclastic rocks from the Pliocene through Pleistocene reflects the increasing maturity of the Cordillera Central volcanic arc through time. Numerous geochemical studies (Dickinson, 1970b; Gill, 1970; Gill and Gorton, 1973) have documented an increase in SiO_2 and K_2O content of volcanic rocks with increasing development of an island arc system. Ragland et al. (1976) and DeBoer et al. (1980) have recently noted the occurrence of calc-alkaline and shoshonitic volcanic belts in western Luzon which also suggests that the arc system has developed to a mature stage of island arc volcanism.

Sandstones

Texture

Numerous studies have documented the distortion of grain size data by the diagenetic alteration of sandstones (Mousinho de Meis and Amadon, 1974; Galloway, 1974; Wilson and Pittman, 1977; Pittman, 1979). To obtain grain size, sieve data reflecting the original size distribution samples were selected that had not been significantly altered by diagenetic processes. The selected samples are all from the Awidon Mesa Formation. Sandstones from the Ilagan Formation are significantly altered and not suitable for sieve analysis.

Grain size statistics (Table 2) indicate that the average Awidon Mesa Formation sandstone is composed of poorly sorted (1.49ø) medium sand (1.85ø). The sorting ranges from moderately sorted (0.69ø) to very poorly sorted (2.37ø), and the mean grain diameter varies from very fine sand (3.39ø) to medium sand (1.35ø). The sandstones are dominantly fine-skewed (0.16) but range from strongly fine-skewed to coarse-skewed. Kurtosis values vary from very platykurtic (0.17) to very leptokurtic (1.86) but are dominantly leptokurtic (1.19).

The grain size data suggest that the sandstones are of fluvial origin. Probability plots (Figure 22a) have characteristics of fluvial channel sands studied by Visher (1969) and point bar data compiled by Glaister and Nelson (1974) (Figure 22b). The saltation population predominates in importance and has a moderately high slope while the suspension population has a lower slope (Glaister and Nelson, 1974). The junction between suspension and saltation populations is sharp and occurs between 2.75 and 3.5ø (Visher, 1969). River sands, like the Awidon Mesa sandstones, are

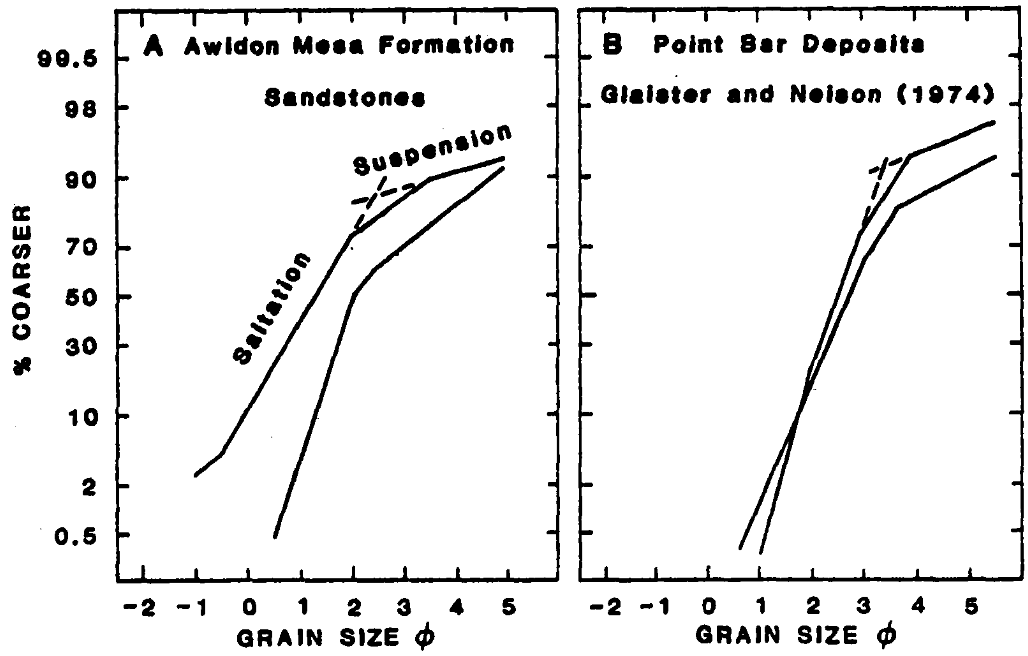


Figure 22. Probability plots of Awidon Mesa sandstones (a) and fluvial sandstones (b)

characteristically leptokurtic (Folk, 1974) and fine-skewed (Friedman, 1967). Most fine-skewed Awidon Mesa sandstones contain more than 5% clay (5.48 to 20.56%) and may be classified (Folk, 1974) as texturally immature. Three samples, however, have less than 5% clay (0.77-4.51%) and are texturally submature. The textural immaturity of the sandstones is consistent with the interpretation that the sediments are of fluvial origin where little effective winnowing of the sediments took place. The submature sandstones which lack clay are similar to clay deficient fluvial sands observed by Moss (1972) and Wilson and Pittman (1977) which were deposited in upper low flow and transitional flow regime currents.

Grain shape and roundness is uniform for all sandstones. The grains are dominantly equant and subangular (Powers, 1953). Euhedral to subhedral feldspars, bipyramidal quartz, and heavy minerals also occur. Volcanic rock fragments are subrounded to rounded indicating preferential rounding of these less stable fragments. The subangular and euhedral grains indicate that the sands have not undergone extensive abrasion during transport from the adjacent volcanic arcs.

Composition

The sandstones are predominantly lithic arkoses and feldspathic litharenites, but litharenites and arkoses are also common. The composition varies significantly between formations as illustrated in Figure 23. The Upper Member of the Ilagan Formation is composed of feldspathic litharenites and litharenites. Two sandstones of the Lower Member (Table 8) which were also studied are litharenites. Feldspars are more abundant in the Awidon Mesa Formation resulting in a greater percentage of lithic arkoses and arkoses.

Color of the sandstones varies significantly and is dependent on composition. The sandstones rich in rock fragments are dark yellowish orange (10YR6/6) to moderate brown (5YR4/4) reflecting weathering (oxidation and hydration) of the rock fragments and associated heavy minerals. Sandstones rich in feldspars, in contrast, are very pale orange (10YR8/2) to light gray (N7) or very light gray (N8). Most of these sandstones are reworked pyroclastic deposits with few rock fragments.

Framework grains, primarily feldspar and rock fragments, are the dominant component of the sandstones, averaging 59% of each sample (Table 8). Strongly zoned andesine is the primary feldspar as recognized by extinction angles of albite twin planes. It increases in abundance from the Ilagan Formation (19%) to the Awidon Mesa Formation (47%). K-feldspar

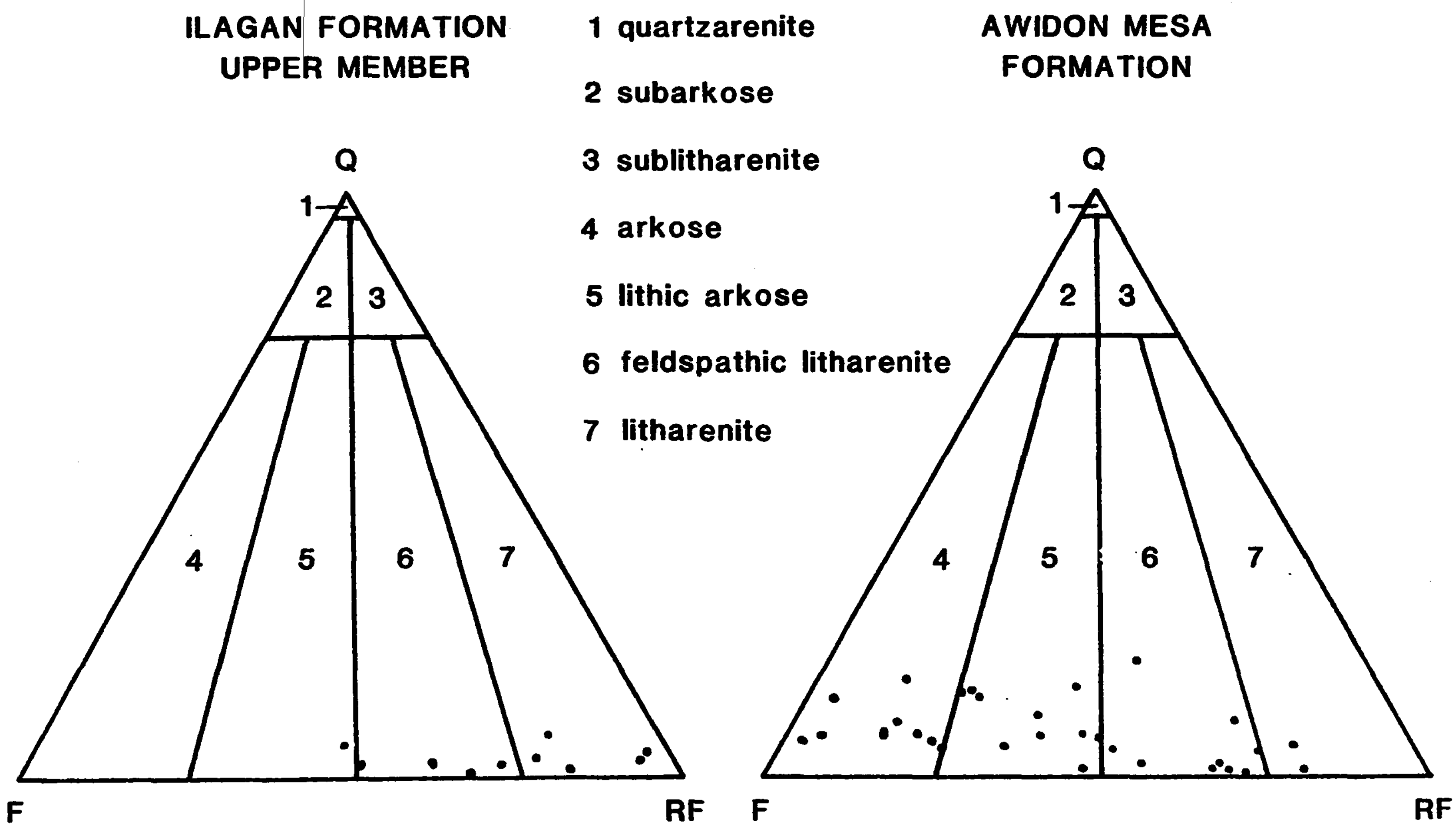


Figure 23. Classification of the sandstones of the Upper Member of the Ilagan Formation and the Awidon Mesa Formation

Table 8. Sandstone composition and mineralogy

Sample no.	Composition: Framework grains	Detrital matrix	Cements: Clay	Calcite	Zeolite	Dissolution features	Porosity
Awidon Mesa Formation							
C3-02	GM[b]						
C4-10	65	18	2	-	-	A	15
C4-31	44	45	-	-	-	S	11
C5-07	60	19	-	-	-	A	21
E1-06	65	1	20	-	-	C	14
E1-09	GM						
E1-18	62	3	22	-	-	C	13
E1A-24	61	18	3	-	-	A	18
E2-01	41	47	-	-	-	S	12
E2-02	69	-	13	-	-	R	18
E3A-02	65	11	5	-	-	S	19
E3A-03	GM						
E4-03	64	5	6	-	-	R	25
E5-35	GM						
E6-02	64	-	3	-	-	-	33
E6-10	GM						
ER1-07a	45	8	24	-	-	A	23
ER1-18b	GM						
ER1-23c	57	3	7	-	11	A	22
ER1-33	GM						
ER3-09	49	29	1	-	-	-	21
M2-03	65	26	-	-	-	R	9
MR1-08	67	10	4	-	-	C	19
P1-114	34	-	26	-	-	A	40
P3-01	58	2	10	-	-	S	30

[a] A=abundant; C=common; S=sparse; R=rare.

[b] Grain mount.

[c] t represents less than 0.5%.

Framework grain mineralogy										
			Rock fragments			Amphiboles		Pyroxenes		
Quartz	K-feldspar	Plagioclase	Volcanic	Sedimentary	Metamorphic, plutonic igneous	Hornblende	Oxyhornblende	Augite	Hypersthene	Opaque minerals
9	-	24	56	5	1	1	-	-	-	4
6	-	45	42	1	t[c]	2	2	-	-	2
13	-	56	18	3	-	7	-	-	-	3
1	-	27	60	1	1	-	4	1	-	5
4	-	48	12	-	3	32	-	-	-	1
6	-	66	15	1	1	7	-	-	-	4
12	-	56	19	-	1	-	-	5	4	3
15	-	45	37	-	1	2	-	-	-	-
10	-	66	3	-	-	10	-	-	-	11
13	-	59	22	1	1	3	-	-	-	1
9	-	49	29	2	1	6	-	-	-	4
6	-	39	38	-	2	8	-	-	-	6
6	-	69	11	t	1	8	-	-	-	5
2	-	30	65	-	1	-	-	2	-	-
3	-	37	40	-	-	2	-	11	6	1
4	-	23	62	5	1	2	-	1	t	2
5	-	75	1	-	-	10	-	-	-	9
5	-	16	64	1	1	6	-	-	-	7
5	-	40	22	5	-	-	1	8	9	10
1	-	17	64	3	-	-	-	9	1	4
8	-	68	14	-	-	4	-	-	-	6
6	-	74	4	-	-	12	-	-	-	4
7	-	47	26	6	10	4	-	-	-	t
2	-	25	63	1	t	-	-	-	-	9
1	-	50	45	-	-	1	-	3	-	-

Table 8. (continued)

	Composition						
			Cements				
Sample no.	Framework grains	Detrital matrix	Clay	Calcite	Zeolite	Dissolution features	Porosity
P4-06	63	1	15	-	-	A	21
P4-31	32	53	4	-	-	R	11
T1-01a	60	-	18	-	-	S	22
T3-02a	59	5	5	-	-	C	31
T3-07a	73	5	2	-	-	S	20
WR1-02a	63	6	13	-	-	C	18
Ilagan Formation							
Upper Member							
E1-26	62	2	7	-	18	A	11
E1-28c	59	3	16	-	-	A	22
E1-35a	GM						
E1-39	GM						
E1-45	67	3	3	-	20	A	7
ER1-43	57	4	3	-	26	A	10
P1-34	63	-	4	-	21	C	12
P1-34a	52	-	-	32	16	S	-
P1-58	60	4	13	-	6	C	17
P1-85	69	-	17	-	-	C	14
Lower Member							
PB1-04	57	-	7	-	18	A	18
PB1-35	59	-	-	-	39	A	2

Framework grain mineralogy										
			Rock fragments			Amphiboles		Pyroxenes		
Quartz	K-feldspar	Plagioclase	Volcanic	Sedimentary	Metamorphic, plutonic igneous	Hornblende	Oxyhornblende	Augite	Hypersthene	Opaque minerals
1	-	27	51	2	1	7	-	3	-	8
11	-	62	16	-	2	4	1	-	-	4
4	-	57	19	-	-	14	-	-	-	6
16	-	70	9	1	2	2	-	-	-	-
4	-	50	27	-	-	2	4	-	-	13
2	-	40	52	-	-	t	-	2	-	4
7	1	15	57	12	5	-	-	t	-	3
-	-	21	44	-	-	2	-	7	-	26
1	-	12	78	4	-	1	-	-	-	4
2	-	25	41	1	-	1	-	14	3	13
2	-	47	43	5	1	-	t	t	-	2
5	1	39	35	4	1	11	-	-	-	4
3	2	4	76	1	6	-	-	7	-	1
4	-	12	59	-	3	-	-	8	8	6
3	-	21	57	8	8	-	t	t	-	2
2	-	25	65	2	-	1	-	2	-	3
1	-	9	86	2	-	-	-	2	-	t
5	-	5	85	2	2	1	-	-	-	t

also occurs in many samples but is rare, usually composing less than 0.5% of the sandstones. The rock fragments are primarily composed of euhedral to subhedral plagioclase laths in a felted microlite groundmass and are interpreted as microlitic volcanic rock fragments (Dickinson, 1970a). The volcanic rock fragments, in contrast to the feldspars, decrease in abundance from the Ilagan Formation (60%) to the Awidon Mesa Formation (32%). Sedimentary, metamorphic, and plutonic igneous rock fragments also occur in the sandstones but are of minor importance.

Quartz is a constant minor component of the sandstones. It increases in abundance from the Ilagan Formation (3%) to the Awidon Mesa Formation (8%) which contains larger water-clear bipyramidal β quartz crystals up to granule size. Many of the Awidon Mesa quartz crystals are embayed or rounded to varying degrees reflecting resorption of the crystals by the original magma. In the Ilagan Formation, the β quartz morphology is not as pronounced and most grains occur as crystal fragments.

Heavy minerals are present in all sandstones but vary significantly in abundance and mineralogy (Table 8). In selected samples of the Lower Member of the Ilagan Formation, heavy minerals are sparse. Highly etched augite or hornblende grains and opaque minerals form only a few percent of each sandstone. In the Upper Member of the Ilagan Formation, heavy minerals are more abundant averaging 12.8% of each sample. Magnetite, ilmenite, augite, and hypersthene are the dominant heavy minerals with minor amounts of green-brown hornblende. Heavy minerals are also abundant in the Awidon Mesa Formation where they constitute 11.5% of the sandstones. Hornblende is the dominant heavy mineral averaging 5% of the sandstones. The opaque minerals, magnetite and ilmenite, are also common forming 4% of the sandstones. Oxyhornblende, augite, and hypersthene occur in some sandstones in variable amounts up to 11%.

The sandstones contain variable amounts of detrital matrix and cement (Table 8). The matrix commonly composes only a few percent of the sandstones. It is locally abundant, however, forming 18 to 53% of 8 samples. Compositionally, the matrix may be classified as protomatrix (Dickinson, 1970a), unrecrystallized clay with minor amounts of silt. Qualitative X-ray diffraction analyses (Table 9) indicate that the matrix clay is either smectite or kaolinite or a smectite-kaolinite mixture as illustrated in Figure 24. Three major types of cement, authigenic clays, zeolites, and

Table 9. Qualitative mineralogy of sandstone clay minerals determined by X-ray diffraction

Sample no.	Dominant clay type[a]	Smectite	Kaolinite
Awidon Mesa Formation			
C4-31	D	X	X
C5-07	D,A	X	
C5-33	D	X	X
E1-09	D	X	
E2-02	A		X
ER1-23c	D	X	
MR1-08	A	X	
P1-114	A	X	
T3-10	D		X
Ilagan Formation			
E1-18	A	X	
E1-45	D	X	
P1-16	A	X	
P1-51a	A	X	
P1-70	A	X	
P1-85	A	X	
P1-111	D	X	X

[a]A=authigenic cement; D=detrital matrix.

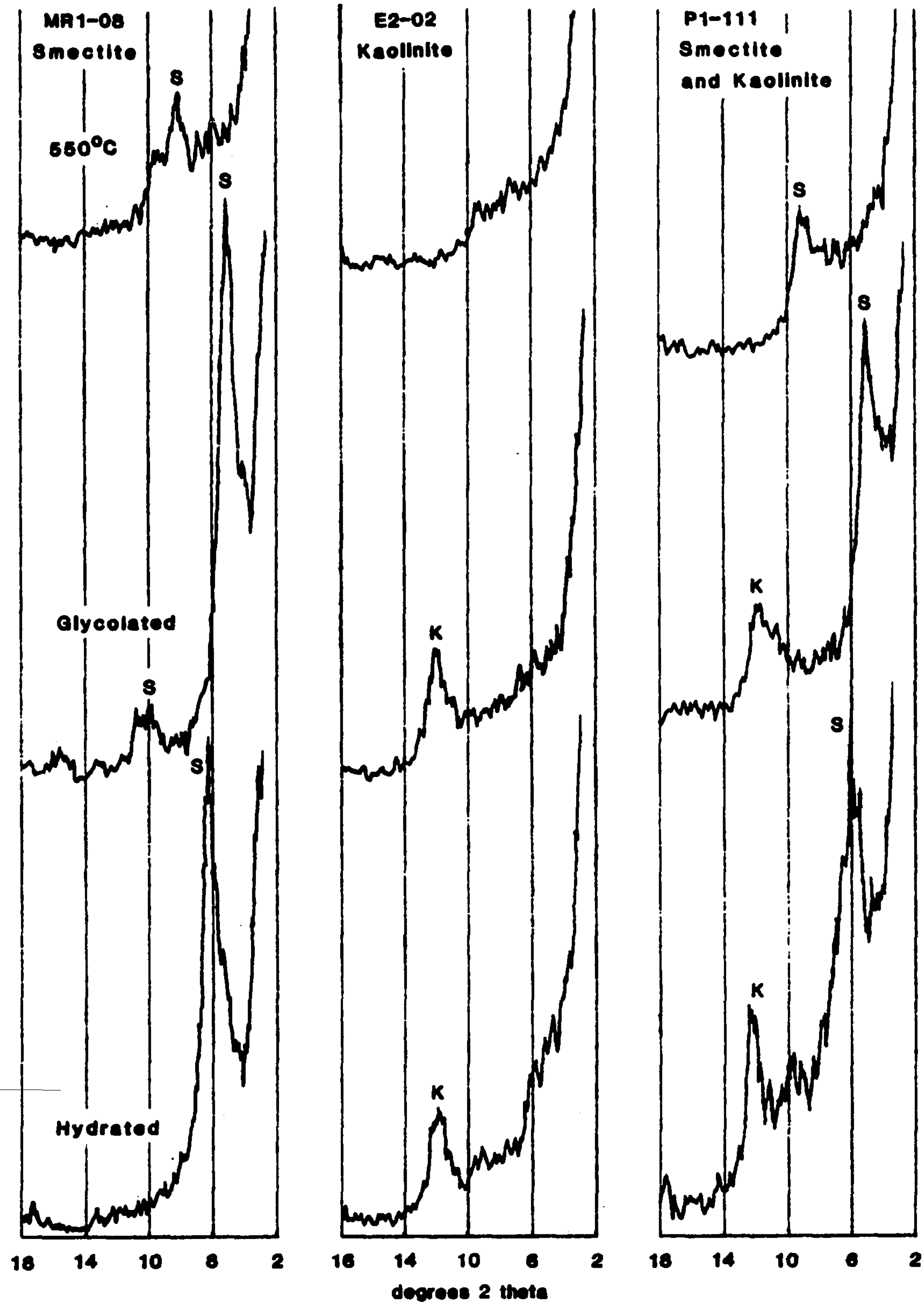


Figure 24. X-ray diffraction tracings of sandstone clay minerals

calcite occur in the sandstones. Minor amounts of iron oxide cement also occur in some samples. The authigenic clays which are differentiated from the matrix clays in Table 9 are dominantly smectite (Figure 24a) or kaolinite (Figure 24b). The characteristics of the various cements are discussed in more detail in the discussion of diagenesis. The cements are more abundant in the Ilagan Formation (26.6%) than the Awidon Mesa Formation (8.5%). Due to the greater amount of cement, the average intergranular porosity of the Ilagan Formation (11.3%) is significantly less than that of the Awidon Mesa Formation (20%).

Provenance To more accurately interpret the heavy mineral variations in the sandstones and provenance, heavy minerals of the fine sand fraction were separated from 10 sandstones and 5 river sands using bromoform and the separation technique outlined by Mathisen (1977). In addition to the hornblende, oxyhornblende, augite, hypersthene, and opaque minerals identified in thin section, sphene and zircon were identified in the heavy mineral concentrates (Table 10). These minerals are rare, comprising less than 3% of the heavy fraction. Numerous grains in the river sands and the Ilagan Formation were classified as composite grains. These are rounded altered grains that contain opaque minerals and heavy mineral crystals thought to be primarily pyroxenes. Several X-ray diffraction powder patterns, however, were more similar to the fayalite pattern suggesting that olivine may also be present.

Distinct variations occur in the abundance of the common heavy minerals which make it possible to differentiate five distinct heavy mineral associations: the pyroxene, green-brown hornblende, oxyhornblende, pyroxene-hornblende, and hornblende-pyroxene associations (Table 11). All

Table 10. Heavy mineral data in number percent for Cagayan basin sandstones and river sands (t represents less than 0.5%)

Sample no.	Augite	Hyper-sthene	Horn-blende	Oxyhorn-blende	Sphene	Zircon	Opaque minerals	Composite
Awidon Mesa Formation								
T3-10	8	-	3	86	1	2	t	-
C4-31	-	-	96	t	t	1	3	-
C5-07	17	1	9	68	t	1	4	-
C5-33	-	-	98	1	t	1	-	-
Ilagan Formation								
P1-111	7	-	6	1	t	t	39	47
P1-12	81	-	11	1	t	2	3	2
P1-34	92	-	2	2	1	1	1	-
P1-51	71	18	4	4	-	-	2	1
P1-70	26	39	21	6	-	-	2	6
E1-45	55	29	5	3	t	t	7	1
Modern River Sands[a]								
MR-01	25	1	16	1	t	1	15	41
MR-02	29	7	6	-	-	3	26	29
MR-03	21	1	39	9	t	3	8	19
MR-04	25	2	37	2	-	3	6	25
MR-05	42	1	36	17	-	-	2	2

[a]Rivers sampled are as follows: MR-01, Cagayan River, Tuguegarao; MR-02, Pinacanauan de Tuguegarao River, Callas; MR-03, Chico River, Tabuk; MR-04, Magat River, Bagabag; MR-05, Aparri beach sand.

Table 11. Heavy mineral associations of the Upper Member of the Ilagan Formation, Awidon Mesa Formation, and Holocene river sands

Associations	Pyroxene (augite-hypersthene)		Hornblende		Oxyhornblende		Pyroxene-hornblende		Hornblende-pyroxene	
No. of samples	5		2		2		3		2	
	$\bar{X}$[a]	σ[b]	$\bar{X}$	σ	$\bar{X}$	σ	$\bar{X}$	σ	$\bar{X}$	σ
Mineralogy										
Augite	65.0	26.2	-	-	12.5	6.4	32.0	9.3	23.0	2.8
Hypersthene	17.2	17.4	-	-	t[c]	-	3.0	3.5	1.5	0.7
Hornblende, green brown	8.6	7.7	97.0	1.4	6.0	4.2	19.3	15.3	38.0	1.4
Oxyhornblende	3.2	1.9	t	-	77.0	13.4	6.0	9.5	5.5	4.9
Opaque minerals	3.0	2.3	1.0	1.4	1.5	2.1	14.3	12.0	7.0	1.4
Sphene	t	-	t	-	t	-	t	-	t	-
Zircon	t	-	1.0	0.0	1.5	0.7	1.3	1.5	3.0	0.0
Composite grains	2.2	2.4	-	-	-	-	24.0	19.9	22.0	4.2

[a] $\bar{X}$=mean.

[b] σ=standard deviation.

[c] t=represents less than 1%.

of the associations are interpreted to be representative of the source area mineralogy. Most amphiboles and pyroxenes are significantly etched by intrastratal solutions (Bramlette, 1929), but thin section analyses indicate that the grains have only been partially dissolved from most samples. In addition to the five associations, a heavy mineral assemblage composed primarily of opaque and composite grains was identified (Table 10). This assemblage is not representative of source area mineralogy and has probably been enriched in equant opaque and opaque-rich composite grains by density and shape sorting processes described by Drummond and Stow (1979) and Stapor (1973).

Three of the heavy mineral associations occur in Plio-Pleistocene sandstones and associated sediments and form distinct mappable petrologic provinces (Suttner, 1974) which correlate with major stratigraphic and structural boundaries (Figure 25). Thin section data of conglomerate matrix, pyroclastic rocks, and sandstones (Tables 5, 6, and 8) indicate that all samples contain primarily pyroxenes, hornblende, or oxyhornblende and may also be classified into the pyroxene, hornblende, or oxyhornblende associations. The pyroxene association occurs throughout the Upper Member of the Ilagan Formation forming a pyroxene province. In contrast, nearly all of the sediments of the Awidon Mesa Formation contain hornblende and constitute a hornblende province. Several pyroxene rich sandstones occur in the Awidon Mesa Formations but are primarily at the base of the formation near the pyroxene hornblende province boundary. The oxyhornblende association also occurs in the Awidon Mesa Formation but is restricted to pyroclastic rocks of the Tabuk plateau where it forms an oxyhornblende province.

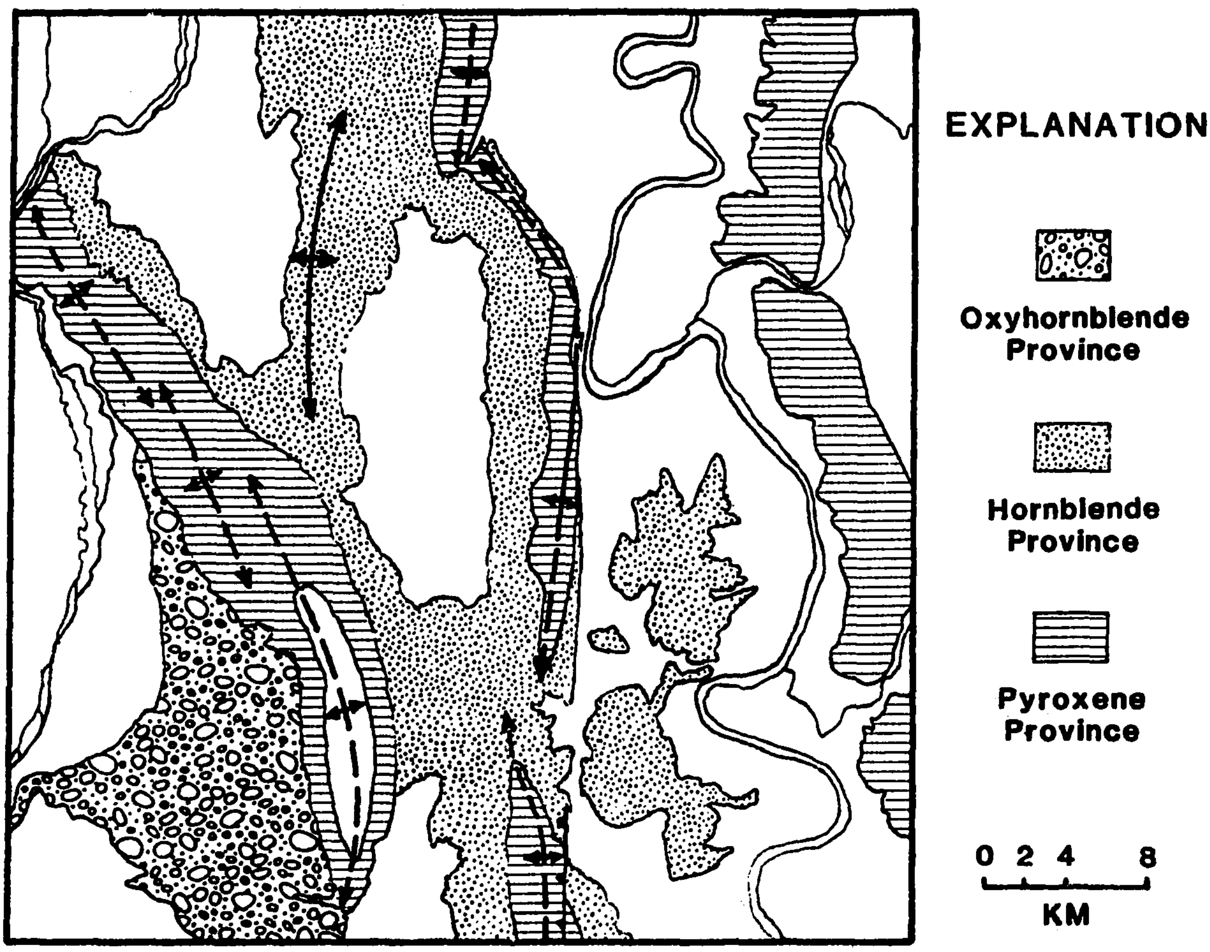


Figure 25. Central Cagayan Valley petrologic provinces

The mineralogy of the river sands records variations in source area mineralogy which can be used to interpret provenance of the Plio-Pleistocene sediments. Sands from the Cagayan and Pinacanauan de Tuguegarao Rivers form the pyroxene-hornblende heavy mineral association (Table 11). The mineralogy of the sand from these rivers, which drain the mountains to the east and south, suggests that pyroxene is the dominant heavy mineral transported to the central Cagayan Valley from the Sierra Madre and the southern part of the basin. The abundance of pyroxene in the pyroxene province of the Ilagan Formation indicates that the Ilagan sediments were derived from the mountains to the east and/or south. Pyroxenes also occur in the hornblende-pyroxene associations (Table 11) of the Chico and Magat rivers which drain the Cordillera Central to the west. This indicates that some sediments of the Ilagan Formation may also be derived from the Cordillera Central. Hornblende is the dominant mineral in the Chico and Magat river hornblende-pyroxene association (Table 10) and reflects a change in the composition of the pyroclastic rocks formed in the Cordillera Central during the Pleistocene. The abundance of hornblende in the hornblende province of the Awidon Mesa Formation, therefore, indicates that sediments of the Awidon Mesa were derived primarily from the Cordillera Central. The oxyhornblende province of the Awidon Mesa Formation indicates that pyroclast flows which originated in the Cordillera Central were oxidized following eruption and deposition on the Tabuk Plateau (Deer et al., 1963).

Other sandstone composition data suggest the same provenance as the heavy mineral data. The volcanic rock fragments of the Ilagan Formation, as well as the pyroxenes, indicate that the Plio-Pleistocene sediments could have been derived from any of the adjacent arcs. The Pleistocene sediments, however, were derived primarily from the Cordillera Central. Aside from the presence of hornblende, this is indicated by the high volcanic quartz and feldspar content of the Awidon Mesa Formation in contrast to the Ilagan Formation. The Awidon Mesa sandstones contain 8% volcanic quartz and 47% plagioclase fledspar while the Ilagan sandstones contain only 3% quartz and 19% plagioclase. The volcanic quartz, plagioclase, and hornblende originated as phenocrysts in the volcanic rocks of the Cordillera Central as a result of increased silicic volcanic activity in the Pleistocene.

Diagenesis

The sandstones of the Ilagan and Awidon Mesa formations have been altered to various degrees by five diagenetic processes. These are: 1) compaction, 2) dissolution of framework grains, 3) formation of authigenic clay, 4) crystallization of authigenic zeolites, and 5) the crystallization of calcite cement.

Compaction Compaction has occurred in all sandstones but is not very extensive. Most grains have floating or point contacts and have been compacted only by grain slippage and rotation (Jonas and McBride, 1977). Most volcanic rock fragments have not been deformed. However, slight ductile deformation (McBride, 1979) of the rock fragments has occurred in some Ilagan sandstones. The lack of compaction features is consistent with the stratigraphic occurrence of the rocks which indicates that they could not have been buried more than about 1,200 m, the maximum possible thickness of the Ilagan and Awidon Mesa Formations. The high porosity, 20-35%, which remained after compaction left the sandstones permeable to fluid migration which resulted in dissolution of unstable grains and precipitation of cements.

Dissolution Partial dissolution of heavy minerals, plagioclase feldspars, and volcanic rock fragments has occurred in most sandstones of the Ilagan and Awidon Mesa formations (Table 8). Dissolution has generally been more extensive, however, in the Ilagan sandstones where more than half the framework grains are often significantly etched. The original form or size of the etched grains is usually indicated by the matrix or authigenic clay rims. Pyroxenes are the only heavy minerals that are extensively etched and nearly always display hacksaw t[illegible]nations as described by Edelman and Doeglas (1931). In most sandstones, the grains have been etched continuously or only once during their diagenetic history. In one sandstone, however, clay rims indicate that the dissolution of pyroxene grains took place in two steps separated by a period of clay formation (Figure 26a). The feldspars are commonly etched along cleavage traces or in a random pattern. Feldspar dissolution has affected both single grains, which are almost completely dissolved in some samples (Figure 26b), and laths in rock fragments (Figure 26c). The dissolution of volcanic rock fragments usually is not complete but produces grains with a distinct amount of microporosity (Figure 26d). In some cases, the dissolution of framework grains has been so extensive the original grains cannot be identified (Figures 26b and e).

The dissolution of plagioclase and volcanic rock fragments is the result of hydration reactions which are ubiquitous in volcanogenic terranes (Surdam and Boles, 1979). The hydration reactions are important because they increase the pH of the interstitial solutions and release cations into solution. This increase in pH and salinity has a direct effect on all subsequent diagenetic reactions. Hay (1966) has noted that plagioclase dissolution occurs in saline, alkaline nonmarine environments, and that the rates of plagioclase dissolution are increased with increasing salinity and pH. The dissolution of plagioclase in the sandstones of the Cagayan Valley probably reflects a similar saline, alkaline chemical environment. The

Figure 26. Photomicrographs of dissolution features of framework grains

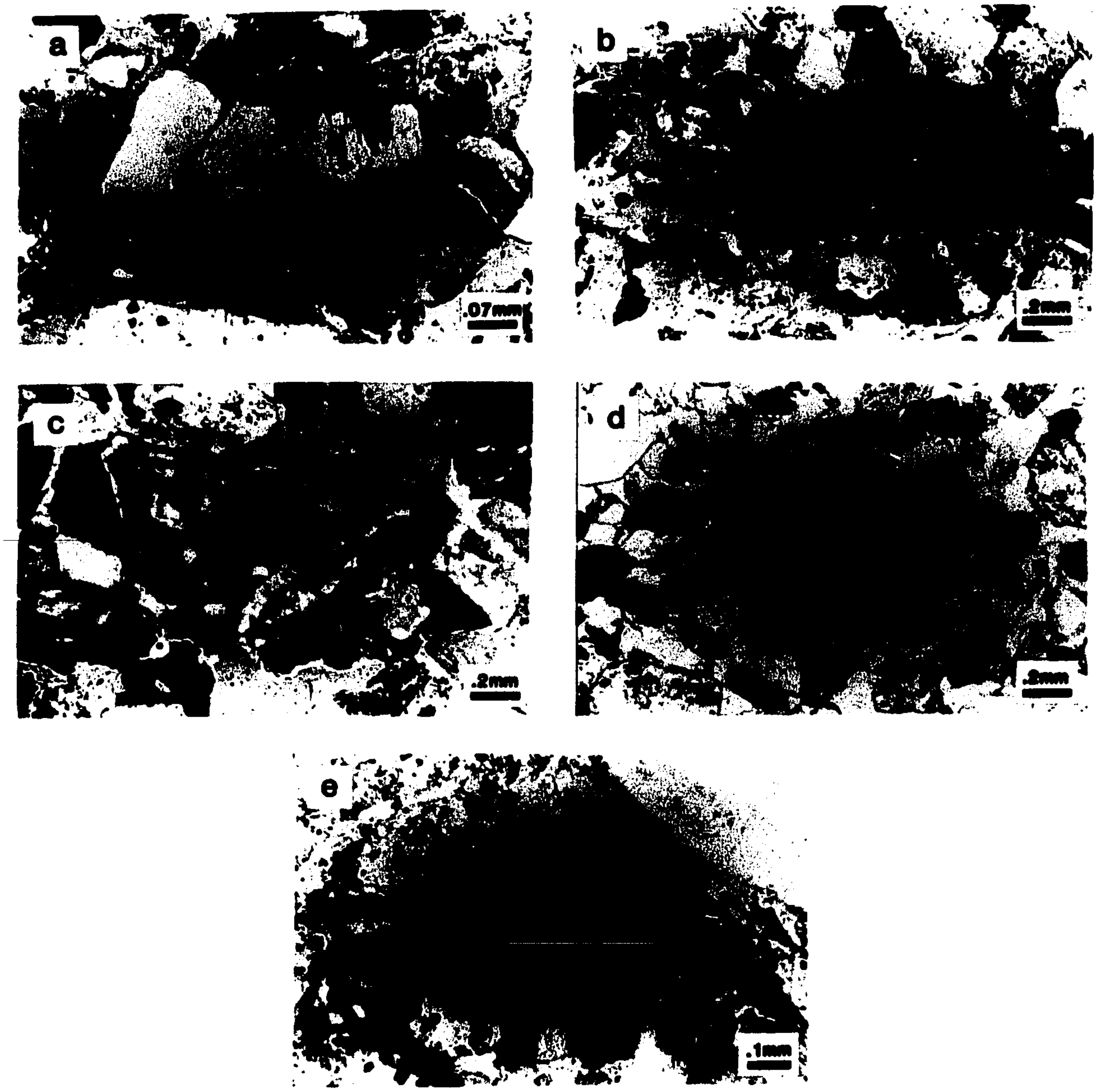
a. Dissolution of augite crystals (p) and formation of hack-saw terminations. A two-step dissolution history is indicated by the authigenic clay rims (c)

b. Extensive dissolution of feldspar grains (f) and other framework grains

c. Dissolution of feldspar laths (f) in rock fragments

d. Partially dissolved volcanic rock fragments (v) with secondary development of microporosity. The zeolite chabazite (c) forms a pore filling cement

e. Pores (p) formed by dissolution of framework grains in a pyroclastic rock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pyroxenes, which are stable under acid conditions (Deer et al., 1963), must have also been etched under the same alkaline conditions.

The dissolution of the various framework grains occurred after initial compaction and has resulted in an increase in porosity for most sandstones. In some samples, the porosity increase has been considerable. If the original sandstone is assumed to have been composed of 60% detrital grains and 40% porosity, nearly half of the framework grains have been dissolved from some samples such as P1-114 (Figure 26b), which contains 34% framework grains, 26% authigenic clay, and 40% porosity. Surdam and Boles (1979) note that large decreases in porosity commonly result in volcanogenic sandstones from the hydration of framework grains and the precipitation of authigenic silicate cements. The conservation or increase of porosity in most Cagayan Valley sandstones indicates that they must have been altered in an open system with a substantial amount of fluid flowing through the sandstones to remove dissolved material.

Authigenic clay Authigenic clays have cemented both the sandstones of the Ilagan and Awidon Mesa formation to varying degrees. The clay averages 7% of the Ilagan sandstones and ranges from 0 to 17% in abundance. In the Awidon Mesa sandstones, the clay averages 8.5% and ranges from 0 to 26%.

The authigenic origin of the clay is interpreted by the identification of clay rims or clay coats which form rinds around detrital sand grains (Galloway, 1974; Wilson and Pittman, 1977). Clay rims consist of clay flakes oriented perpendicular to the surface of the detrital grain (Galloway, 1974). Smectite commonly forms homogeneous clay rims of uniform distribution and thickness (up to 20 μ) in the Ilagan and Awidon Mesa sandstones as illustrated in Figure 27a and c. The smectite was identified from optical properties, X-ray patterns, and morphology of the clay as viewed with the SEM. Figure 27d illustrates the typical honeycomb-like appearance of authigenic smectite as described by Wilson and Pittman (1977). Clay coats, which show preferred orientation parallel to the grain surfaces (Galloway, 1974), also commonly occur in the Ilagan and Awidon Mesa sandstones. The clay coats are commonly stratified and of variable thickness as illustrated in Figure 27e. Smectite is the dominant clay mineral that forms clay coats and occurs as layers of flat slightly irregular flakes (Figure 27f) as illustrated by Scholle (1979). The clay coats in the Plio-Pleistocene sandstones appear to have been formed by illuviation of colloidal material onto the grains as described by Galloway (1974).

X-ray diffraction (Figure 24) analyses of authigenic clay minerals from 9 sandstones (Table 9) indicates that both smectite and kaolinite have formed authigenically. Smectite (Figure 24a and 27a) is the more common clay cement while kaolinite (Figure 24b and 27b), which occurs primarily in the Awidon Mesa Formation, is of minor importance.

The authigenic clay began to form early in the diagenetic history of the sandstones. Significant amounts of clay had coated detrital grains in many samples before significant dissolution of the framework grains had occurred as indicated by the relict clay coats in many samples such as Figure 26a. In numerous samples, the clay continued to accumulate until the pores were partially or completely filled (Figure 27g and h).

<u>Zeolites</u> Zeolites are important cementing minerals in the Ilagan Formation. Nearly every Ilagan sandstone examined contained zeolites which average 16% but form up to 39% of the samples (Table 8). Zeolites are also

Figure 27. Photomicrographs of sandstone authigenic clays

a. Authigenic smectite clay rims(s) of uniform thickness and distribution

b. Authigenic kaolinite (k)

c. SEM photomicrograph of smectite clay rim(s) overlying grain surface

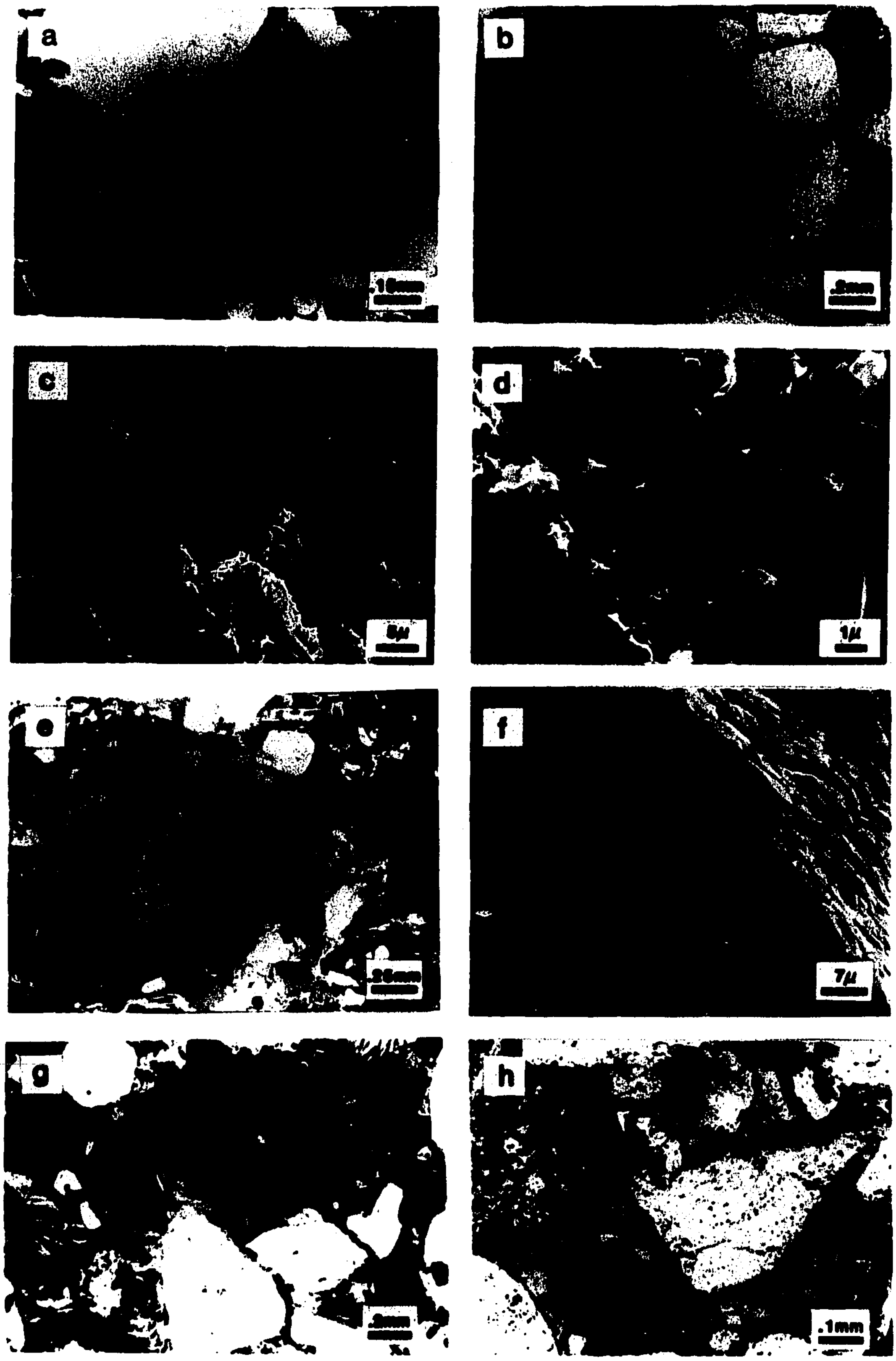
d. SEM photomicrograph of smectite clay rim honeycomb-like morphology

e. Stratified clay coats(s) of variable thickness composed of smectite

f. SEM photomicrograph of authigenic clay coat formed by oriented smectite flakes

g. Smectite clay coats(s) which partially and completely fill original pores

h. Sandstone that has been almost completely cemented by clay(s) which probably formed as clay rims which eventually filled pores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abundant in the Ilagan Formation tuffs (Table 6). In the Awidon Mesa Formation, zeolites were observable with the petrographic microscope in only one sample. SEM analyses, however, suggest that submicroscopic crystals may be forming in other Awidon Mesa sandstones. The Ilagan zeolites primarily occur as pore lining crystals which average 50 μ in length (Figure 28) but also form pore filling cements (Figure 26d). Clinoptilolite, stilbite, analcime, and chabazite were identified by X-ray diffraction tracings (Figure 29) and crystal morphology (Figure 28).

The distribution of the zeolite mineral species varies stratigraphically in the Ilagan Formation. Clinoptilolite and stilbite are the principal zeolites of the Upper Member but some crystals of chabazite were also observed (Figure 28b). These zeolites usually occur in the same rock associated with authigenic smectite as indicated by petrographic microscope observations, X-ray diffraction patterns (Figure 29c), and SEM analyses (Figure 28c and d). In most samples, clinoptilolite and stilbite occur as isolated pore lining crystals (Figure 28a), but they also form continuous coatings (Figure 28g and h). In one lithic tuff (P1-51a, Table 6), clinoptilolite and stilbite totally cement the sample (Figure 20f) reducing the porosity to 2%. Analyses of three sandstones indicate that the Lower Member contains pore lining analcime (Figure 29a and Figure 28e and f) and pore filling chabazite (Figure 29b and Figure 26d) in contrast to the Upper Member. Additional analyses by Erik Kvale (Iowa State University, personal communication, 1981) indicate that analcime and chabazite are the principal zeolites of the Lower Member.

Zeolites are common authigenic minerals in volcaniclastic sandstones which form by the interaction of saline, alkaline aqueous solutions, and

Figure 28. Photomicrographs of authigenic pore lining zeolites from the Ilagan Formation

a. Euhedral pore lining crystals of clinoptilolite and stilbite

b. SEM photomicrograph of pore lining crystals of clinoptilolite (c) and stilbite (s) with interpenetrant twinning and chabazite cubes (ch)

c. SEM photomicrograph of pore lining monoclinic clinoptilolite and stilbite crystals

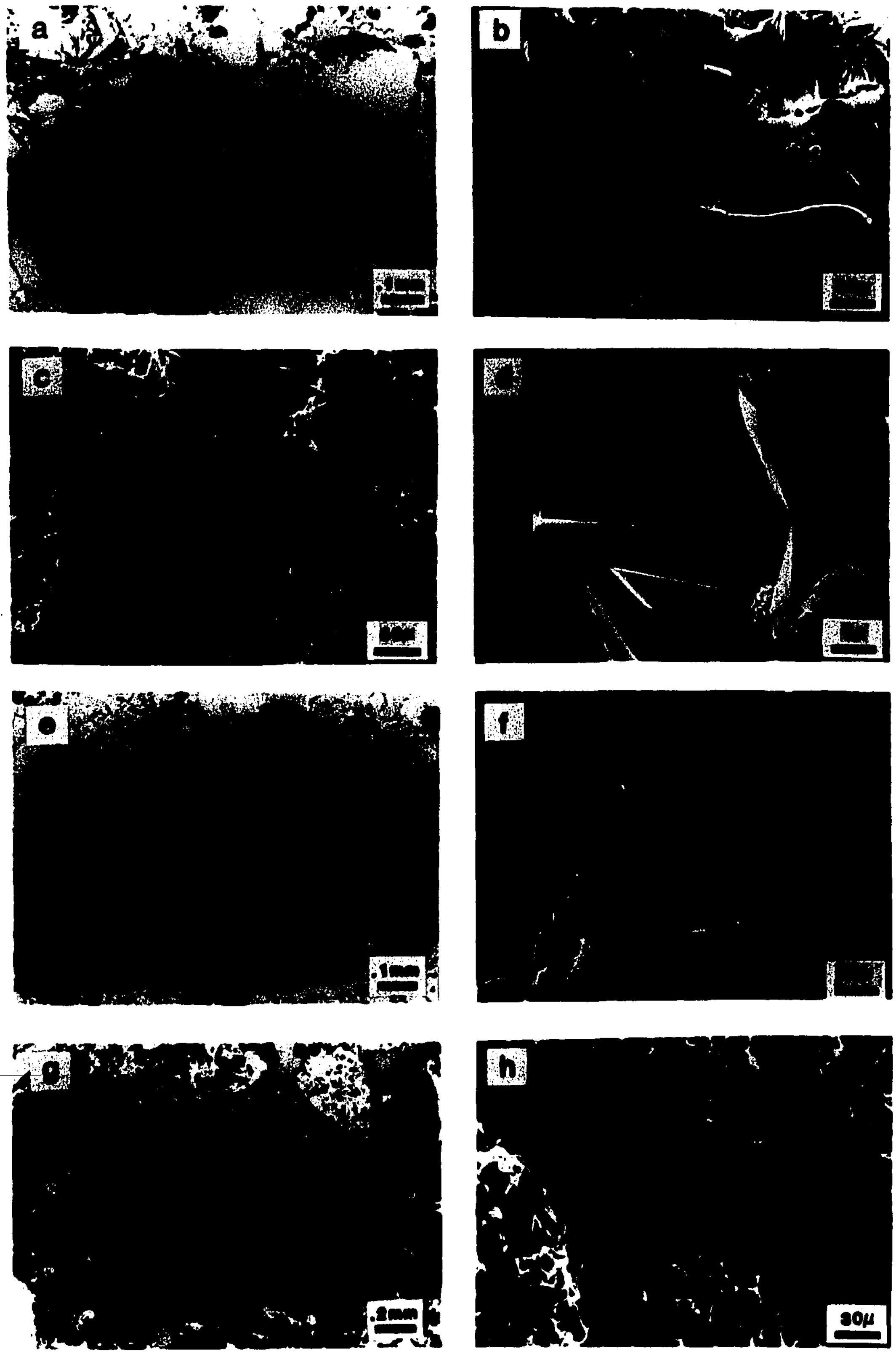
d. SEM photomicrograph of pore lining clinoptilolite with authigenic smectite clay rims forming between crystals

e. Pore lining analcime crystals in typical cubo-octahedral form

f. SEM micrograph of pore lining analcime crystals displaying the typical cubo-octahedral form

g. Zeolite coats formed by clinoptilolite and stilbite

h. SEM photomicrograph of clinoptilolite and zeolite crystals forming continuous coatings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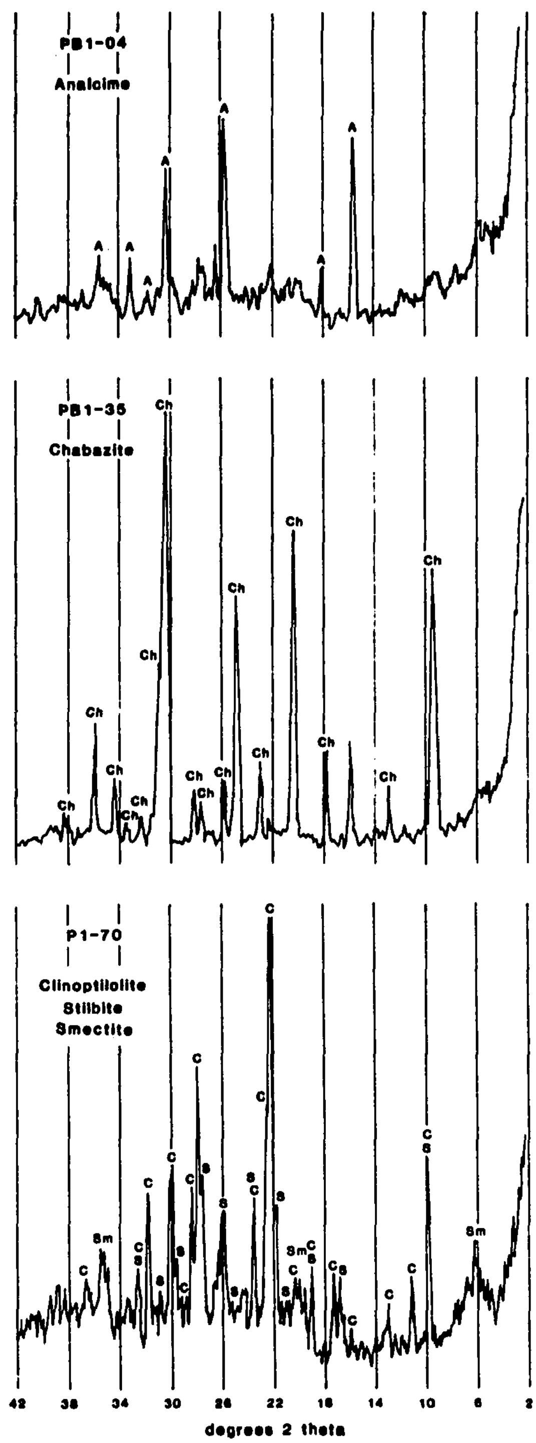


Figure 29. X-ray diffraction tracings of zeolite cements

unstable volcanic detritus (Deffeyes, 1959; Hay, 1966; Surdam and Boles, 1979). Volcanic glass and plagioclase commonly dissolve in saline, alkaline solutions, and contribute Si and Al to the pore waters. The zeolines then precipitate from solution. Plagioclase feldspars are abundant in the Ilagan Formation and have been significantly dissolved in most samples. The zeolite reaction:plagioclase → zeolite is, therefore, a major diagenetic reaction that has taken place in the Ilagan sediments. The zeolitic reaction:glass → zeolite may also have taken place, but there is no petrographic evidence that glass shards were present and dissolved from the Ilagan Formation. The association of smectite clay rims with the zeolites may indicate that glass was present. This is because the alteration of glass to smectite is an important factor in providing the chemical environment, particularly an alkaline pH and high activity of silica, suitable to the formation of zeolites (Hay, 1966). The volcanic origin of the Ilagan sediments and the occurrence of volcanic glass in the overlying Awidon Mesa Formation also suggest that glass may have been present in the Ilagan Formation. If the glass was present, dissolution and the reaction: glass → zeolite has occurred, and no relict shard textures were preserved.

The Ilagan Formation sandstones may be divided into two diagenetic facies, the clinoptilolite-stilbite facies which occurs throughout the Upper Member and the analcime-chabazite facies which occurs at greater depth in the Lower Member. Clinoptilolite, stilbite, and chabazite commonly form at shallow depths under low temperatures. Analcime, in contrast, forms by the alteration of low temperature zeolites such as clinoptilolite and is stable at higher temperatures and greater burial depths (Hay, 1966). The alteration of low temperature zeolites to analcime with increased temperature and depth of burial has been attributed to: 1) burial metamorphic reactions due to increased temperature and pressure (Aoyagi and Kazama, 1980); 2) increases in salinity and pH (Sheppard and Gude, 1969); 3) kinetic factors where analcime forms later than zeolites such as clinoptilolite (Sheppard and Gude, 1969), and 4) various combinations of these factors (Hay, 1966; Coombs and Whetten, 1967). The Ilagan Formation zeolite facies have been subjected to different burial depths, temperatures, and pressures as well as different reaction times. These factors have definitely contributed to the differentiation of the analcime-chabazite facies. Increases in salanity and pH may also have been important factors, but these cannot be assessed at this time.

Calcite Sparry calcite is a localized cement which forms concretions in the Ilagan Formation. The calcite dominantly fills pores that are lined with authigenic clay and zeolites but also partially replaces some plagioclase grains. The formation of calcite cement is a common diagenetic reaction in volcaniclastic sediments (Galloway, 1974). Surdam and Boles (1979) note that the initial hydration reactions of volcaniclastic rocks are sources of Ca^{++} and that organic matter is one of the major sources of HCO_3^-. They suggest that the amount and distribution of organic matter in sediments is probably the limiting factor relative to carbonization reactions. The Ilagan concretions, therefore, indicate that a significant amount of organic matter was deposited with the sediments. Calcite concretions were then formed after decay of the organic matter and the formation of HCO_3^-.

Diagenetic sequence The Ilagan and Awidon Mesa sandstones display diagenetic features indicative of both mechanical and chemical diagenesis.

The distribution of these features with depth is summarized in Figure 30. Compaction is the main mechanical event and began shortly after deposition with a limited amount of grain slippage and rotation. Chemical diagenesis also began shortly after deposition and has had the greatest effect on the sandstones. Authigenic clay rims and coats began to form first at shallow depths, several hundred meters, and continued to form with burial to depths of approximately 1,000 m. Dissolution of pyroxenes, plagioclase, and volcanic rock fragments also began to occur at shallow depths. The relict clay rims and coats which indicate the original grain form suggest that significant amounts of authigenic clay had formed before the grains were extensively dissolved. The relict rims and coats also indicate that significant amounts of additional compaction did not occur with depth. Extensive dissolution of framework grains occurred at relatively shallow depths and has also been extensive at greater depths where grains are dissolved after zeolite pore-filling cement has formed. Pore-lining zeolites of the clinoptilolite-stilbite facies began to form in sandstones buried several hundred meters and are abundant in samples to a depth of approximately 900 m. Below this depth to approximately 1,200 m, the analcime-chabazite zeolite facies occurs reflecting a greater depth of burial, longer reaction time, and possible geochemical variations of pore fluids. The last diagenetic event in some sandstones was the precipitation of pore filling sparry calcite. The calcite has formed concretions at depths of 400 to 1,200 m cementing sandstones which already contained pore lining clays and zeolites.

The diagenetic sequence just described, which is based on outcrop samples and maximum possible burial depths, is interpreted to reflect

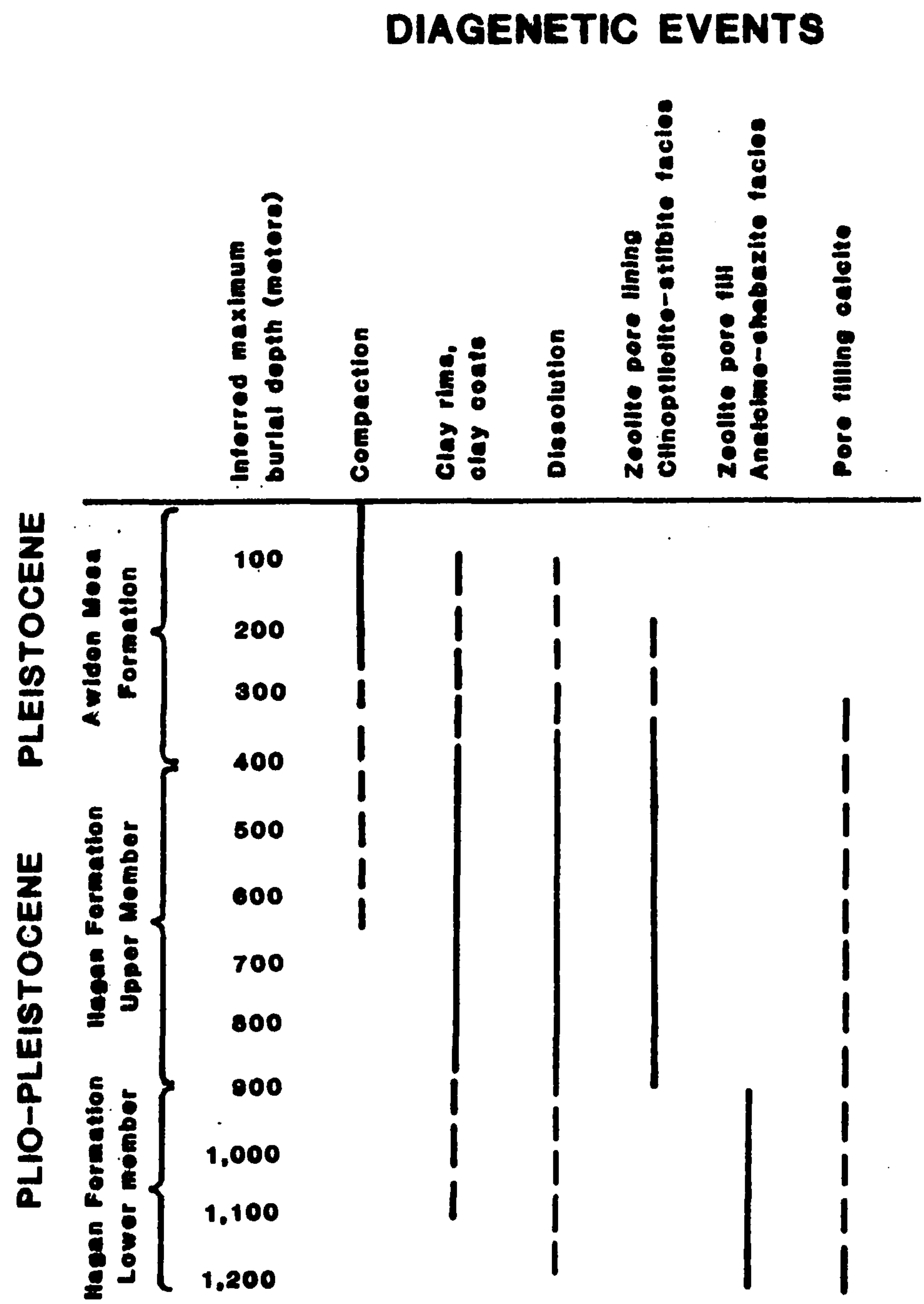


Figure 30. Diagenesis of the Ilagan and Awidon Mesa sandstones. (The vertical bars indicate the interpreted depth range of major diagenetic events)

diagenetic change with increasing depth of burial and age of the deposit in contrast to alteration by weathering processes. Alteration during weathering may have occurred before deposition and/or after folding and erosion of the anticlines. Alteration probably occurred during weathering in the source area, but this cannot be distinguished from the extensive diagenetic alteration. The weathering would have affected all samples equally because the climate did not change significantly in Northern Luzon during the Plio-Pleistocene. The increase in the alteration of the Ilagan Formation sandstones in contrast to the Awidon Mesa Formation sandstones must, therefore, reflect diagenetic alteration. Alteration during weathering of the sampled outcrops may also have occurred but cannot be differentiated from the diagenetic alteration. Both the Ilagan and Awidom Mesa formations were weathered for the same period of time following erosion of the anticlines. Alteration of the samples during outcrop weathering should be similar for both formations if the alteration occurred during weathering. The more extensive alteration of the Ilagan Formation cannot be attributed to weathering and must, therefore, reflect increased diagenetic alteration with burial depth and time. Diagenetic interpretations of outcrop samples have also been made by Galloway (1974) who did not notice any significant alternation of the outcrop samples by weathering when compared to core samples.

The diagenetic features and sequence of the Ilagan and Awidon Mesa sandstones reflect a phase of burial diagenesis typical of volcaniclastic sandstones (Figure 31). This phase essentially consists of compaction and the formation of authigenic clay rims and coats (Stage 2) and then phyllosilicate and/or zeolite pore filling cement (Stage 3) as described by

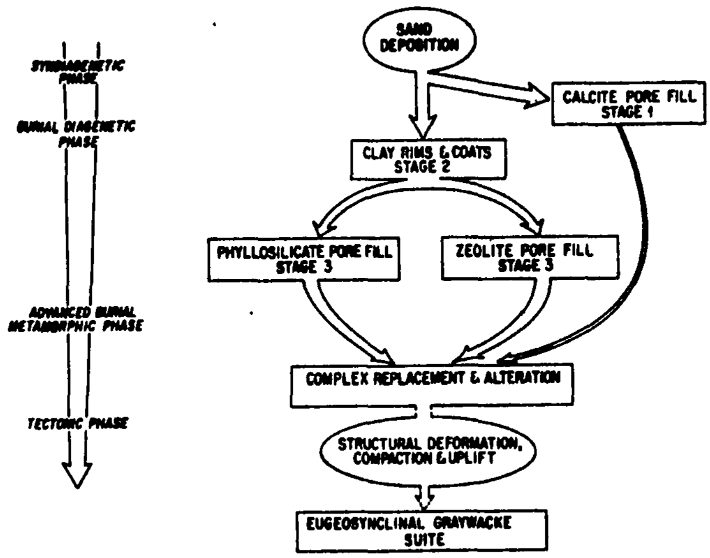


Figure 31. Diagenetic phases and corresponding chemical diagenetic stages in volcaniclastic sandstones (from Galloway, 1974)

Galloway (1974). Calcite pore-filling cements (Stage 1) commonly form first in some volcaniclastic sandstones (Galloway, 1974) but did not occur early in the Ilagan and Awidon Mesa formations. This probably reflects the transitional marine and fluvial environments of the Ilagan and Awidon Mesa sandstones which contained less Ca^{++} and HCO_3^- than volcaniclastic rocks deposited in marine environments. Significant amounts of calcite have formed concretions, however, during the later stages of diagenesis, Stages 2 and 3. The lack of complex replacement and alteration features indicates that the Ilagan and Awidon Mesa sandstones have not been subjected to the advanced burial metamorphic phase defined by Galloway (1974).

Mudrocks

Texture

Mudrocks of the Upper Member of the Ilagan Formation and the Awidon Mesa Formation are primarily claystones and mudstones (Figure 32) as

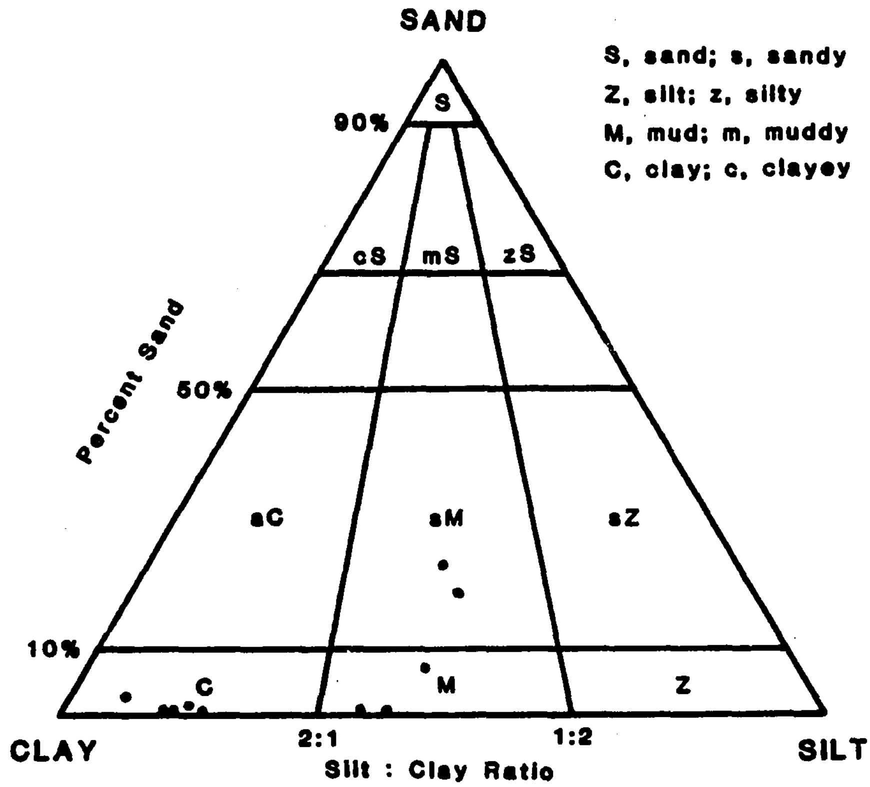


Figure 32. Textural classification of Cagayan Valley mudrocks (after Folk, 1954)

defined by Folk (1954) and Blatt, Middleton, and Murray (1980). Clay content of the mudrocks averages about 73%. Sandy claystones and mudstones also occur in the formations and contain up to 22% very fine to coarse sand. Sand, silt, and clay percentages range from 0.4 to 22%, 16-44%, and 44-92.5%, respectively. Statistical parameters of the mudrocks sampled are recorded in Table 12. The average mean grain size of the mudrocks is 9.72Ø, clay, and the average standard deviation is 2.88Ø, very poorly sorted. The very fine grain size and very poor sorting suggest that the mudrocks were deposited from suspension.

Table 12. Statistical parameters of selected mudrocks

Sample no.	Grain mean	Inclusive graphic standard deviation
Awidon Mesa Formation		
C1-09	8.08	4.30
C3-34	10.36	2.71
E1-11	8.30	4.70
E8-03	10.73	2.20
ER1-22	10.05	1.86
P4-11	9.02	2.70
Ilagan Formation Upper Member		
E1-23b	9.58	2.84
E1-41	11.70	1.84
P1-23a	8.72	3.10
P1-48a	10.65	2.61

Composition

The mudrocks which are dominantly light olive gray (5Y5/2) in color are composed primarily of clay minerals, quartz, and feldspar (Table 13). Smectite is the most abundant clay mineral and occurs in nearly all samples studied. Kaolinite also occurs in many samples but, from the relative intensity of the X-ray peaks, is not as abundant as smectite. It is also more common in the Awidon Mesa Formation than in the Ilagan Formation. The clay minerals form three different types of clay mineral assemblages. These assemblages, as illustrated by X-ray diffraction tracings (Figure 33), are 1) a smectite assemblage, 2) a kaolinite assemblage, and 3) a smectite-kaolinite assemblage. The variations in clay mineral composition do not, in general, correlate with color variations of the mudrocks.

A variety of minerals occurs in the sand fraction of the mudrocks. Plagioclase is the dominant mineral in the sand while bipyramidal volcanic

Table 13. Qualitative mineralogy of mudrock samples

Sample no.	Smectite	Kaolinite	Quartz	Feldspar
Awidon Mesa Formation				
C1-19	X	X	X	
C4A-03	X	X	X	X
E6A-06a	X		X	X
E8-03	X	X	X	X
E8-08a	X	X		
ER1-19	X		X	X
ER1-26	X		X	X
MR1-07		X	X	X
P1-113	X	X	X	
P2-04		X	X	X
P3-10	X		X	
P4-02	X		X	X
P4-29	X	X	X	X
SM1-07	X		X	X
Ilagan Formation				
E1-23b	X		X	X
E1-32	X		X	X
E1-38b	X	X	X	X
P1-23a	X		X	X
P1-48a	X		X	X
P1-87	X		X	X
P1-99a	X	X	X	X

quartz is a constant minor constituent in each sample. Other minor constituents are hornblende, augite, magnetite, zircon, and occasional glass shards. The sandy claystone with the greatest sand fraction, E1-11, contains very poorly sorted sand composed of volcanic quartz, plagioclase, hornblende, and zircon and lacks epiclastic detritus. Several volcanic quartz granules also occur in the sand. The poor sorting and mineralogy suggest that sandy claystones with volcanic quartz granules such as E1-11 may have formed from the *in situ* alteration of fine grained pyroclastic deposits.

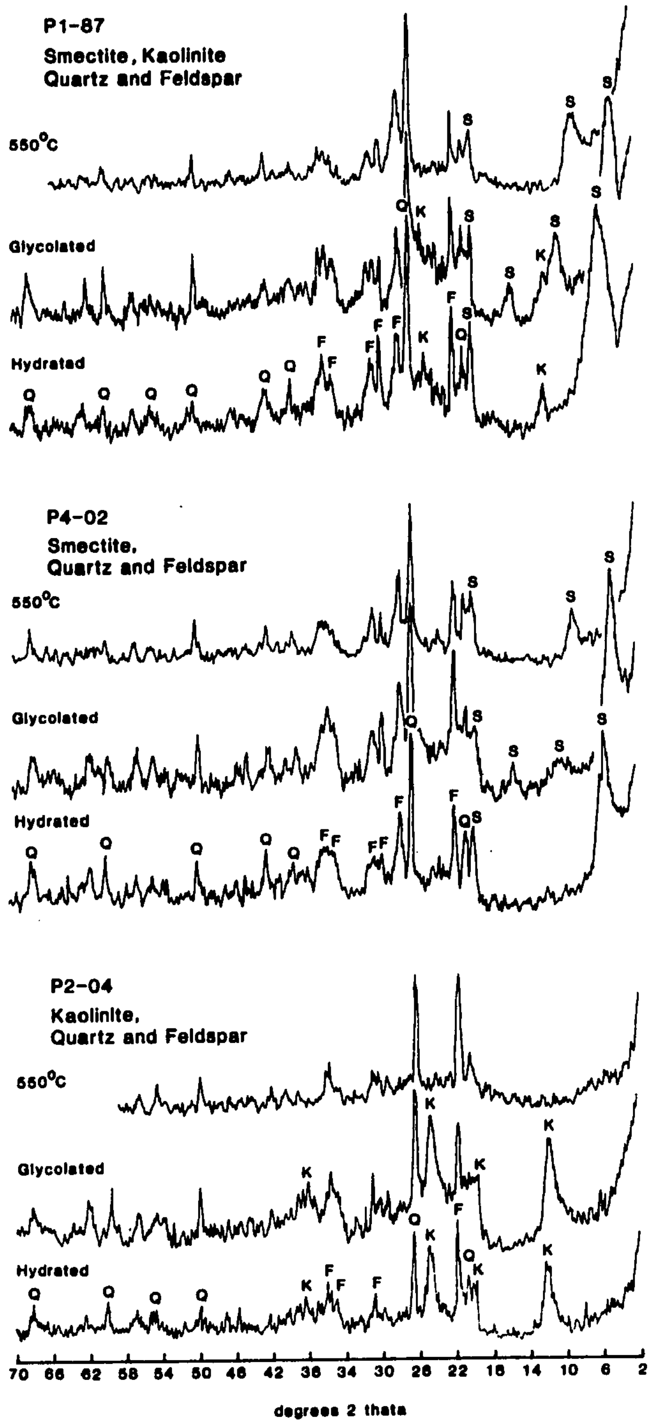


Figure 33. X-ray diffraction tracings of mudrocks

The tectonic setting and the stratigraphic association with pyroclastic deposits also suggest that the mudrocks were formed by the alteration of fine grained volcaniclastic and pyroclastic material. Keller (1970) and Nagasawa (1978) have noted that smectites and kaolinite both commonly form as a result of the weathering of pyroclastic deposits. At low temperatures and pressures, kaolinite tends to form under nonalkaline conditions while smectite forms under alkaline conditions. The abundance of smectite suggests that alkaline conditions have prevailed in the mudrocks and that most may be classified as bentonites, i.e. smectite rich clays that have formed from the <u>in situ</u> alteration of volcanic ash.

Paleosols occur in numerous mudrocks reflecting various stages of soil development. Horizons with weakly developed blocky structure were provisionally interpreted as weakly developed paleosols in the field. More detailed analyses of the micromorphology, clay distribution, and soil chemistry are needed to more accurately assess the degree of soil development and its significance. Horizons characterized by iron enrichment and iron rich pisolites are interpreted to be more highly developed paleosols. These deposits are mottled from pale olive (10Y6/2) to moderate reddish brown (10R4/6), moderate red (5R4/6) and light brown (5YR5/6). The horizons have a weakly developed structure formed by fine to medium subangular blocky peds. The iron pisolites which are up to granule size are well developed with sharp boundries. Concentric layers of authigenic smectite are associated with some pisolites in thin section and are interpreted as argillans. Ferrans and opaque cutans interpreted as sequans are also present. Further studies of these deposits are needed to more accurately assess the degree of soil development, but iron enrichment, cutans and the weak structure indicate that a significant amount of soil development took place during the Plio-Pleistocene. The accumulation of iron probably reflects development of a plinthite horizon which developed on an older more stable part of the landscape which underwent seasonal fluctuation of the water table (Buol et al., 1973).

STRUCTURE AND TECTONIC HISTORY

The central part of the Cagayan Valley may be divided into three major structural provinces (Figure 34). Along the western side of the valley is a very strongly folded belt (Caagusan, 1978) which has been referred to as the Kalinga foothills (Durkee and Pederson, 1961). It is characterized by broad synclines and steeply folded anticlines which have been uplifted and eroded, exposing Miocene turbidites of the Mabaca River Group. A homoclinal belt is present along the eastern margin of the valley. Distinct homoclinal ridges, dipping from 10 to 14 degrees to the west, are formed by the resistant Ilagan and Callao formations. A moderately folded belt (Caagusan, 1978) trends north-south throughout the central part of the Cagayan Valley and is here referred to as the Cagayan Valley anticlinal belt as named by Durkee and Pederson (1961). This belt contains about twenty major asymmetrical doubly plunging anticlines which are separated by broad, flat, alluviated synclinal valleys. The Cabalwan, Pangul, and Enrile anticlines occur in the central part of the belt where most of the field work for this study was concentrated. Extensive exposures of the Ilagan and Awidon Mesa formations occur along the flanks of these anticlines which vary from about 15 to 25 kilometers in length. The Pangul and Enrile anticlines are overturned on the east flank similar to several other anticlines in the belt (Durkee and Pederson, 1961; Christian, 1964). The Cabalwan Anticline is also asymmetrical but has a steeper dip of 38 degrees along the western flank and more gentle dips of 7 to 10 degrees along the east flank.

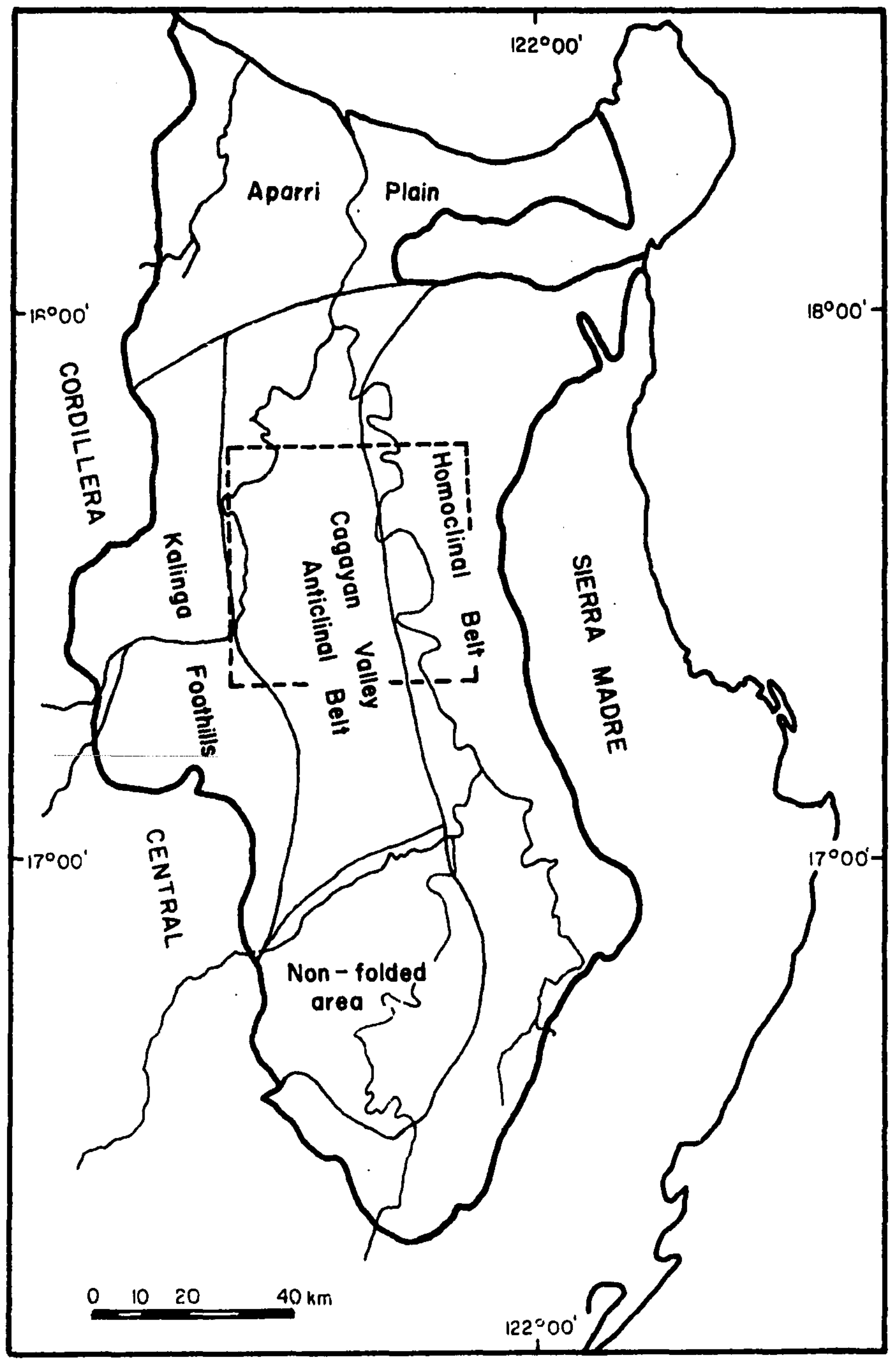


Figure 34. Cagayan Valley structural provinces

The Plio-Pleistocene tectonic history of the Cagayan basin was dominated by regional uplift with very little tilting or compressional deformation (Christian, 1964). This resulted in the transition from marine to terrestrial sedimentation. A greater rate of uplift and volcanic activity took place in the Cordillera Central. Three thousand m of displacement has occurred along a major north-south trending late Miocene to Pleistocene fault zone in the Kalinga foothills (Caagusan, 1980). Uplift of the Cordillera is also reflected in the development of asymmetry in the basin, which became steeper to the west, and the migration of the basin axis approximately 10 km to the east during the Plio-Pleistocene (Christian, 1964; DeBoer et al., 1980). Middle Pleistocene oversteepening of the Cordillera Central resulted in mass gravity failure of sediments blanketing the Cordillera. Décollement occurred, and the sediments slid off the east ~~flank of the~~ Cordillera forming the strongly asymmetrical to overturned anticlines of the Cagayan anticlinal belt (Figure 35). The gravity sliding mechanism was proposed by Christian (1964) because seismic data indicate that the folds do not persist with depth, and Plio-Pleistocene compressive features such as thrust faults and folds are not known in the Cordillera Central foothills or the Sierra Madre. The homoclinal ridges bordering the Sierre Madre were probably formed in the middle to late Pleistocene and may reflect renewed uplift of the Sierra Madre in response to reactivation of the Quezon trench to the east.

A review of the mechanics of gravity sliding supports the interpretation (Christian, 1964) that gravity sliding is the principal cause of folding in the Cagayan Valley. Viscosity variations within a sequence of sediments and the thickness of the beds are two of the more important

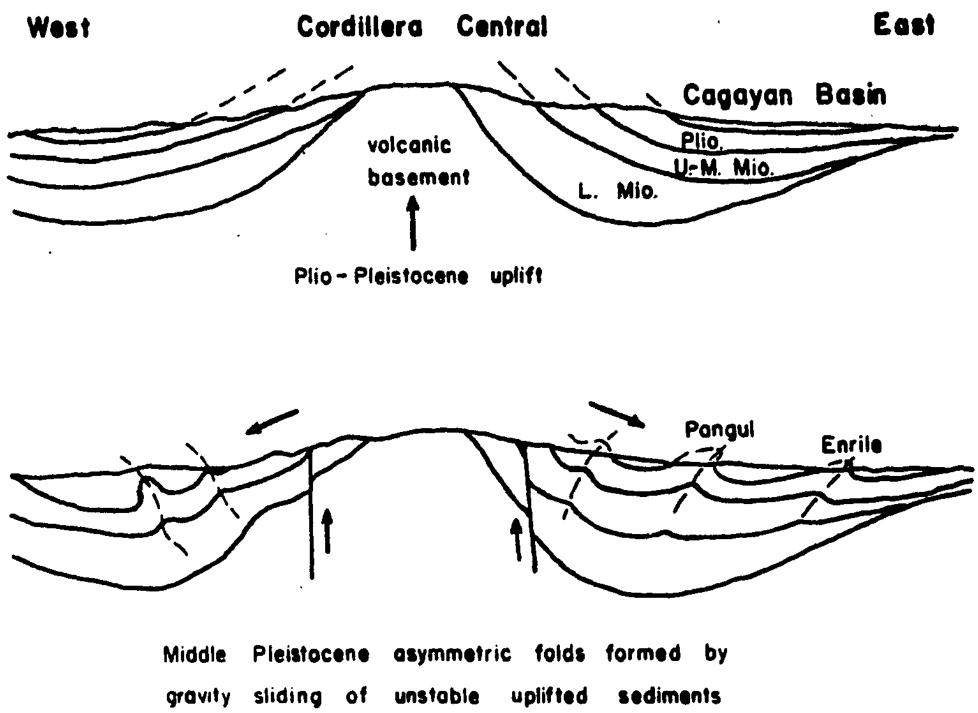


Figure 35. The formation of folds by Middle Pleistocene gravity sliding (after Christian, 1964)

factors which control the development of a décollement zone and the velocity of the slide (Kehle, 1970). In the Cagayan Valley, the shales of the Mabaca River Group form a 8,000 meter thick low viscosity zone where décollement could occur. Kehle (1970) notes that the contribution of gravity increases with 1) the dip of the potential slide mass because the component of gravity in the direction of motion increases with dip, and 2) the depth of burial of the potential décollement zone because the shear stress in a dipping bed increases linearly with depth of burial. As the Cordillera Central was uplifted, the dip of the Plio-Pleistocene terrestrial sediments began to steepen over an already thick and deeply buried potential décollement zone. The slope at which gravity sliding may begin depends on the fluid pressure (Hubbert and Rubey, 1959; Lemoine, 1973) and the weight or thickness of the slide mass (Goguel, 1948). Various calculations have been worked out indicating that a 2 to 4 km thick layer will flow on slopes ranging from 2 to 5 degrees (Goguel, 1948; Hubbert and Rubey, 1959). Hose and Danes (1973) have noted that once sliding is initiated, it may create a mass deficiency in the hinterland and a mass surplus along the leading edge. The net effect in the rear would be uplift in the autochthon to provide a slight local gradient increase. Other factors which may contribute to a potential gravity slide are changing temperatures which may cause variations in pore pressure (Lemoine, 1973) and earthquakes (Pierce, 1973) which may help to trigger a slide. Along the Cordillera Central, higher temperatures and earthquakes would have been significant because of the volcanic and tectonic activity associated with the active island arc. Altogether, conditions in the Cagayan Valley were favorable to initiate gravity sliding of the Plio-Pleistocene fluvial sediments during the Pleistocene. The asymmetric to overturned folds occurring in the valley today are the fold type to be expected as a result of gravity sliding.

A firm date for the Pleistocene folding can be obtained from radiometric dates of pyroclastic deposits in the Awidon Mesa Formation. Pyroclastic deposits occur in an intraformational unconformity along the western flank of Pangul Anticline. Tuffaceous sediments of the lower part of the Awidon Mesa Formation are folded as part of the Anticline while younger flat lying sediments of the Tabuk plateau overlie them unconformably. Radiometric dating of tuffs or pumice cobbles in this sequence will indicate the time span during which the folding took place.

GEOMORPHOLOGY

The central Cagayan Valley is about 40 kilometers wide and contains three major rivers. The largest, the Cagayan River, is a large meandering stream which drains the southern Cagayan Valley and flows to the sea in the north. The other two rivers, the Chico and Pinacanauan de Tuguegarao, are braided tributary streams. The Chico River drains the Cordillera Central which rises to an elevation of 2,216 m west of the valley, and the Pinacanauan de Tuguegarao River drains the Sierra Madre which rises to 1,833 m to the east.

Four major geomorphic areas may be recognized in the central Cagayan Valley. From west to east these are: 1) the Tabuk plateau, 2) the Cagayan anticlinal belt, 3) the Cagayan River plain, and 4) the homoclinal belt.

The Tabuk plateau is a Pleistocene alluvial fan and terrace complex at elevations between 170 and 300 meters along the mountain front (Figure 36a). It extends for 15 kilometers between the mountain front and Pangul Anticline and is underlain by a complex of pyroclastic deposits, fluvial conglomerates, and sandstones which unconformably overlie folded Miocene and Pliocene strata. The alluvial fan, which has a one degree slope along the southern part of the plateau, has been dissected by the braided Chico River in the north (Figure 36a and b). Here a sequence of four major rock cut terraces are well-preserved between Pangul Anticline and the mountain front. Following folding of the anticlines, the east flowing Chico River was diverted to the north as a result of ponding by Pangul Anticline. At present there is a sharp right angle bend in the Chico River Valley where it changes from a consequent to a subsequent valley.

Figure 36. Photographs of geomorphological features, central Cagayan Valley

a. Tabuk plateau with remnant of alluvial fan extending from mountain front at 1 degree slope

b. Braided Chico River as viewed from upper terrace level

c. Pangul anticline viewed from Espinosa Ranch

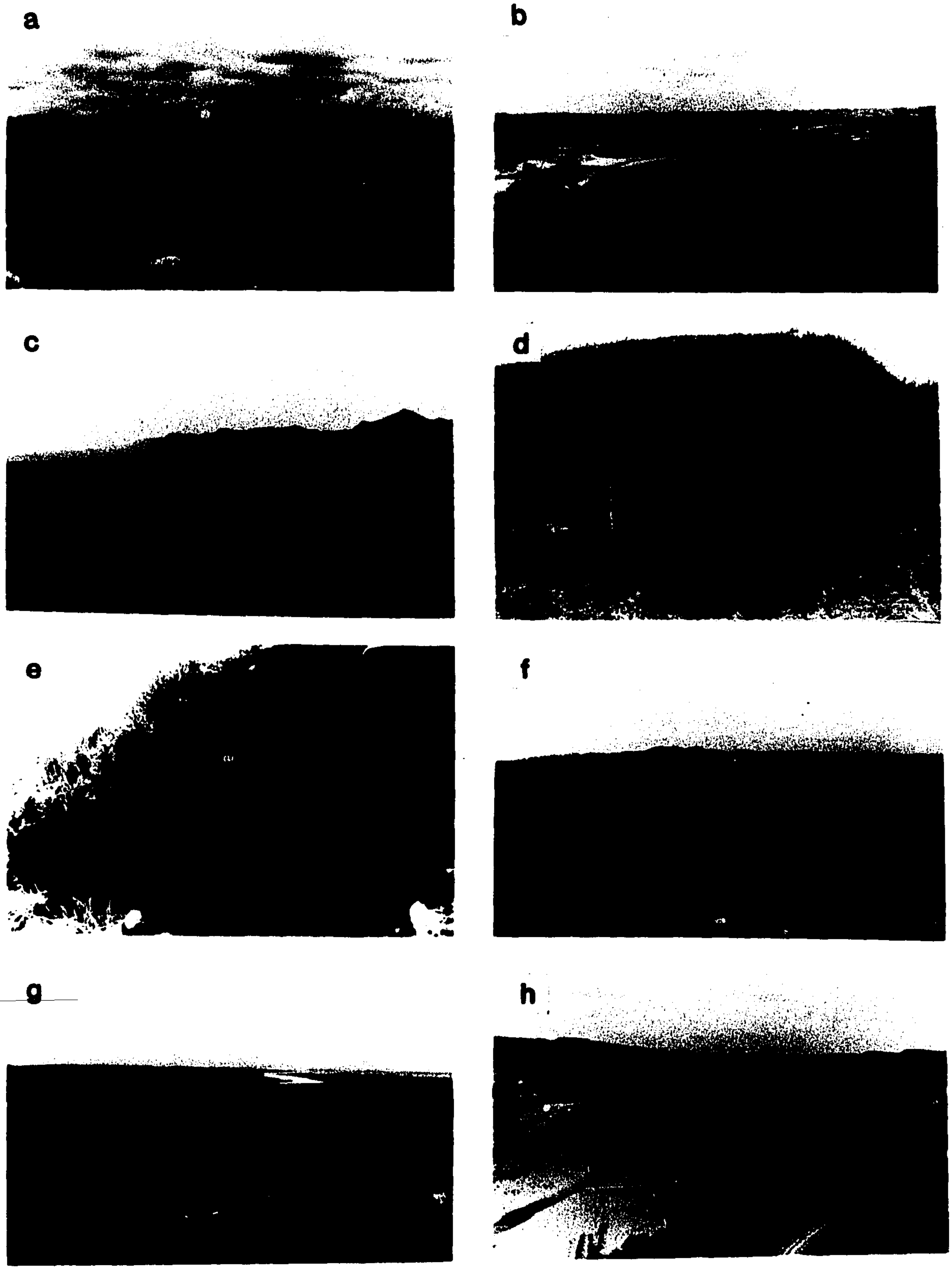
d. Pebble band 1 m beneath surface, southern Pangul Anticline

e. Pebble to cobble conglomerate which occurs in the Awidon Mesa Formation, northern Cabalwan Anticline

f. Stripped surface developed on pebble to cobble conglomerate, Madrigal Ranch, southern Cabalwan Anticline

g. Cagayan River and point bar south of Tuguegarao

h. The braided Pinacanauan de Tuguegarao River which flows between homoclinal ridges formed by the Ilagan and Callao formations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The Cagayan anticlinal belt is characterized by dissected anticlines and synclines of varying relief. They are surficially expressed as linear trends of grass covered rugged hills and ridges or escarpments separated by broad, flat, alluviated synclinal valleys which constitute good grazing land and areas for rice production (Durkee and Pederson, 1961). Pangul Anticline, a large breached anticline with 350 meters of relief, has the greatest relief of any of the anticlines in the area (Figure 36c). In contrast, Cabalwan Anticline has the lowest relief, about 100 meters. Wasson and Cochrane (1979) suggest that dissection of the anticlinal hills is not far advanced. They note, however, that headward extension of the drainage lines has reached the axes of all anticlines. These drainage lines are primarily consequent, but important subsequent valleys do occur. Homoclinal ridges, upheld by resistant pyroclastic deposits, conglomerates, and sandstones, have been formed by the subsequent streams and are extensive along the flanks of the anticlines.

Several different erosional processes are operating on the flanks of the anticlines. Slope wash and soil creep are the two most important processes (Wasson and Cochrane, 1979). Surficial layers are being transported down slopes as indicated by imbricated pebble and cobble bands as much as a meter beneath the surface at several localities (Figure 36d). Slumping also occurs and is most extensive in interbedded thin sandstones and mudstones of the Lower Member of the Ilagan Formation exposed in Pangul Anticline.

Many of the National Museum archaeological sites and fossil localities in the anticlinal belt occur in or are associated with 10 to 80 cm thick discontinuous gravel beds previously thought to be terrace deposits (Lopez, 1972). Detailed stratigraphical studies of the anticlines indicate that the gravels are lag gravels derived from a conglomerate or conglomeritic interval in the upper part of the Awidon Mesa Formation. Along the northern part of Cabalwan Anticline, a massive 5.5 m thick pebble-cobble conglomerate is present within the Awidon Mesa Formation (Figure 36e). At the southern part of the anticline, stripped (terrace-like) surfaces are developed on this conglomerate (Figure 36f) which is more resistant to erosion than the surrounding finer grained sediments. Along the southern nose of the Anticline, erosion has been more extensive resulting in lag gravel covered hills.

The Cagayan River plain is a flat, synclinal, alluviated plain 10 to 20 m above sea level between the anticlinal and homoclinal belts. The Cagayan River meanders across the plain eroding along cutbanks and depositing sediment on large point bars, such as the bar just southwest of Tuguegarao which is up to 1.5 km in radius (Figure 36g). Several terrace levels are well-developed along the eastern side of the Cagayan Valley. Durkee and Pederson (1961) have mapped the Isabela Syncline along the same general trend as the river, and Wasson and Cochrane (1979) have noted that it is, therefore, a consequent stream.

Two major cuestas form the homoclinal belt along the east side of the Cagayan Valley. These are upheld by the Callao Formation and the Lower Member of the Ilagan Formation. The Ilagan Formation which dips to the west at about 10 degrees forms a steep east-facing, north-south trending escarpment with 120 m of relief along the Pinacanauan de Tuguegarao River (Figure 36h). The river flows east of the escarpment in a strike valley developed in the underlying, less resistant shales of the Baliwag Formation before it cuts through the cuesta at Peñablanca and joins the Cagayan River. Slumps are common along the escarpment which consists of thinly bedded sandstones and mudstones. Farther to the east the Callao Limestone forms the most prominent cuesta and escarpment with up to 400 meters of relief. Karst topography is well-developed on the 10 degree dip slope where archaeologists of the National Museum have located numerous cave sites. The geomorphology of these sites has recently been briefly described by Wasson and Cochrane (1979). One major stream, the Pinacanauan de Tuguegarao River, cuts across the Callao cuesta forming the Callao canyon. This river which parallels strike for a short distance between the two ridges has been referred to as a consequent stream by Wasson and Cochrane (1979).

PALEONTOLOGY

The first mammal fossils recovered from the Pleistocene strata of the Cagayan Valley were rhinocerous teeth. The teeth, found in 1936 near Tabuk, were described by von Koenigswald (1956) who noted that they were not identical with the species from China, Taiwan, or Indonesia. He assigned the teeth to a new species, *Rhinocerous philippinensis nov. sp.*, and suggested that it was an endemic species which will only be better known when more and better material is available.

In the early 1970s, poorly preserved remains of a middle Pleistocene vertebrate fauna were found in the central Cagayan Valley during archaeological investigations (Fox, 1971; Fox and Peralta, 1974). Teeth are well-represented in the fossil collections and allow the preliminary identification of the elephants, *Stegodon* and *Elephas*, and rhinocerous, carabao (Bovidae), pig, and crocodile. The recovery of antlers and fragments of carapace indicate the presence of deer and turtle. All the finds so far are disarticulated, and the larger fossils are abraded indicating they were transported by streams before burial. From a brief description of the morphology of the elephant teeth, Lopez (1971) suggested that two forms of *Stegodon* were represented by the finds near Solana, Cagayan. Due to a lack of comparative material, he did not assign them to a specific species. Maglio (personal communication with Dr. Richard Shutler, Jr., project archaeologist, 1976) examined one of the *Elephas* molars from Solana and classified it as *Elephas cf. maximus* but noted that it has a similar structure to the extinct *Elephas hysudrindicus*.

During the 1978 and 1979 field seasons, 32 vertebrate localities, which contained primarily teeth and bone fragments, were found along both flanks of Cabalwan, the northwest and northeast flanks of Enrile, and the northwest flank of Pangul anticlines. The stratigraphic occurrence of these fossil localities is noted on graphic sections of the Awidon Mesa Formation (Figures 10 and 11). All vertebrate localities known at this time occur stratigraphically beneath the Liwan Pyroclastic Complex. Most fossils were found as lag on the Awidon Mesa Formation, but 5 localities were discovered where fossils occur _in situ_. One bone fragment was found _in situ_ in the conglomerate that, when eroded, produces tektite bearing lag gravels. The association of this vertebrate fossil and tektites suggests that the fossil is the same age as the tektite, 0.92 ± 0.17 Myr B.P. This association also explains why tektites and fossils are commonly found together at many National Museum sites where lag gravels are present.

The most significant fossil discovery during this investigation was a small _in situ_ elephant skull, probably _Stegodon_, with two complete upper molars. This was found in an indurated sandstone along the northeast flank of Pangul Anticline (Figure 37a and b). A number of bone fragments and a worn pig tooth were recovered on the surface in this area which can be radiometrically dated since tuffs occur a few meters above and below the elephant skull. Stone tools occur on the surface at this site, but none, as of yet, have been recovered _in situ_. One cobble tool, composed of basalt, which was found on the surface appears to be ancient because all surfaces are differentially weathered.

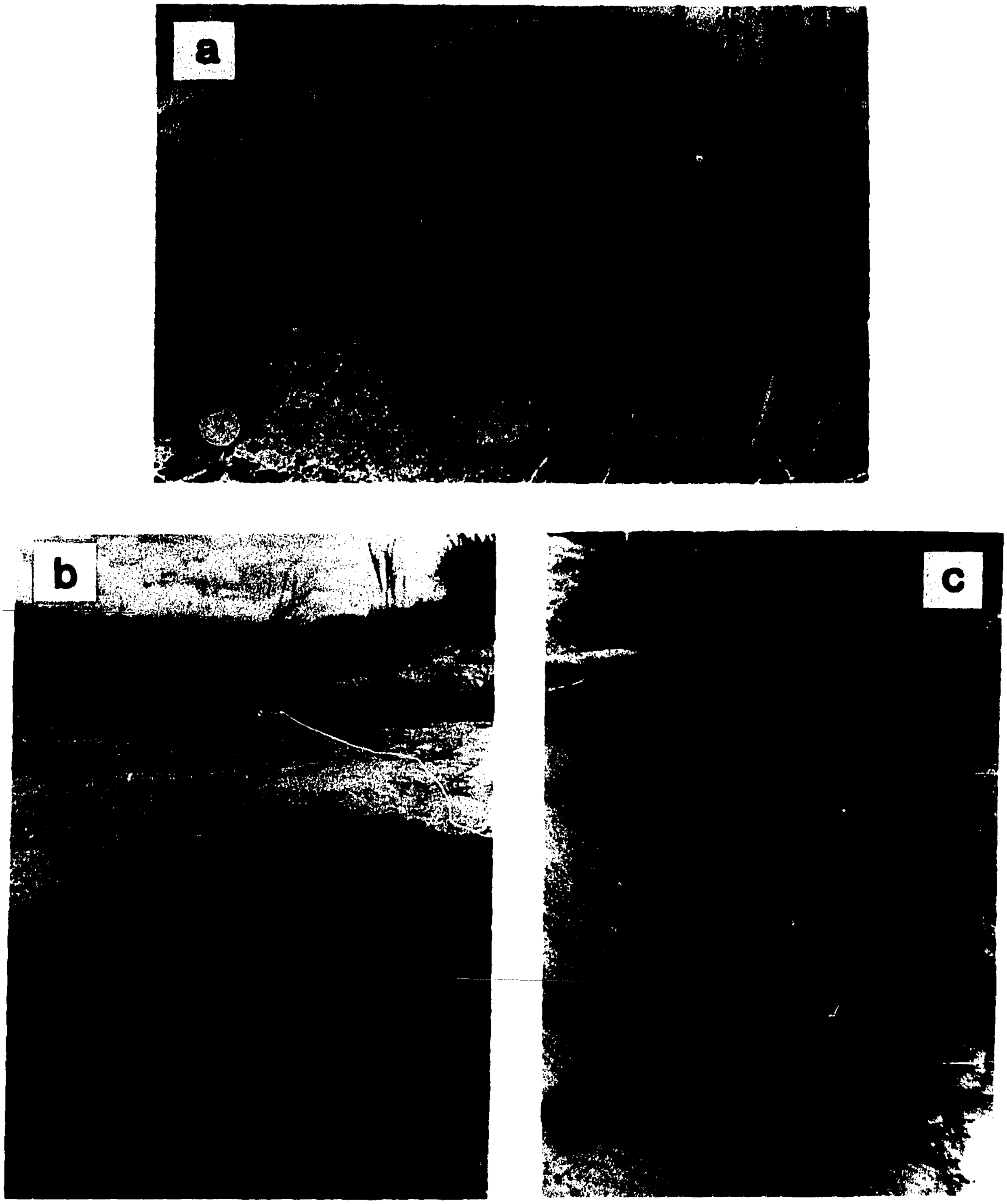
In addition to the elephant locality, four other fossil sites were discovered where vertebrates occur _in situ_. At one locality along the

Figure 37. Photographs of *in situ* vertebrate fossils

a. Stegodon skull with two complete upper molars eroding out of indurated sandstone

b. Stegodon skull eroding from upper part of trough cross bedded sandstone

c. Bovide bone (identified from associated teeth and canon bones) in clay ball in base of trough cross bedded sandstone


a
b
c

western flank of Cabalwan Anticline, carabao (Bovidae) teeth, vertebrae, and canon bones were scattered on the surface and traced to a sandstone outcrop where a tooth and several bones were recovered *in situ*. The fossils occurred in cobble size clay balls in the basal large scale trough cross bedded sandstone of a point bar sequence (Figure 37c) that is overlain by a radiometrically datable tuff-breccia.

Plant fossils are present in the Ilagan Formation as well as the Awidon Mesa. Tree molds and permineralized wood are common along with leaf impressions. Fern leaves are particularly well-preserved and abundant in some tuffs. A group of 8 vertical molds in a tuff-breccia of the Tabuk Pyroclastics probably represents a group of trees that were buried by a pyroclast flow. The molds are from 7 to 28 cm in diameter, extend for up to 4 m into the deposit, and are equally spaced.

Pleistocene vertebrate fossils have been found throughout the Philippines (Beyer, 1956; von Koenigswald, 1956) indicating that land-bridges once connected the Philippines with mainland Asia. It has been suggested that land-bridges once existed between Luzon, Taiwan, and Mainland China and between the southern Philippines and the islands of Indonesia on the Sunda shelf (Figure 38) (Dickerson, 1928; DeTerra, 1943; Movius, 1949; von Koenigswald, 1956; Sartono, 1973). Land-bridges must have existed at times when sea level was lowered during the Pleistocene thus allowing large vertebrates to migrate to the Philippines. The exact migration route cannot be stated with certainty at this time because tectonic activity has broken any possible bridge.

Research to date suggests that the Cagayan Valley vertebrate fauna migrated to Luzon from Taiwan. Von Koenigswald (1956) favors this

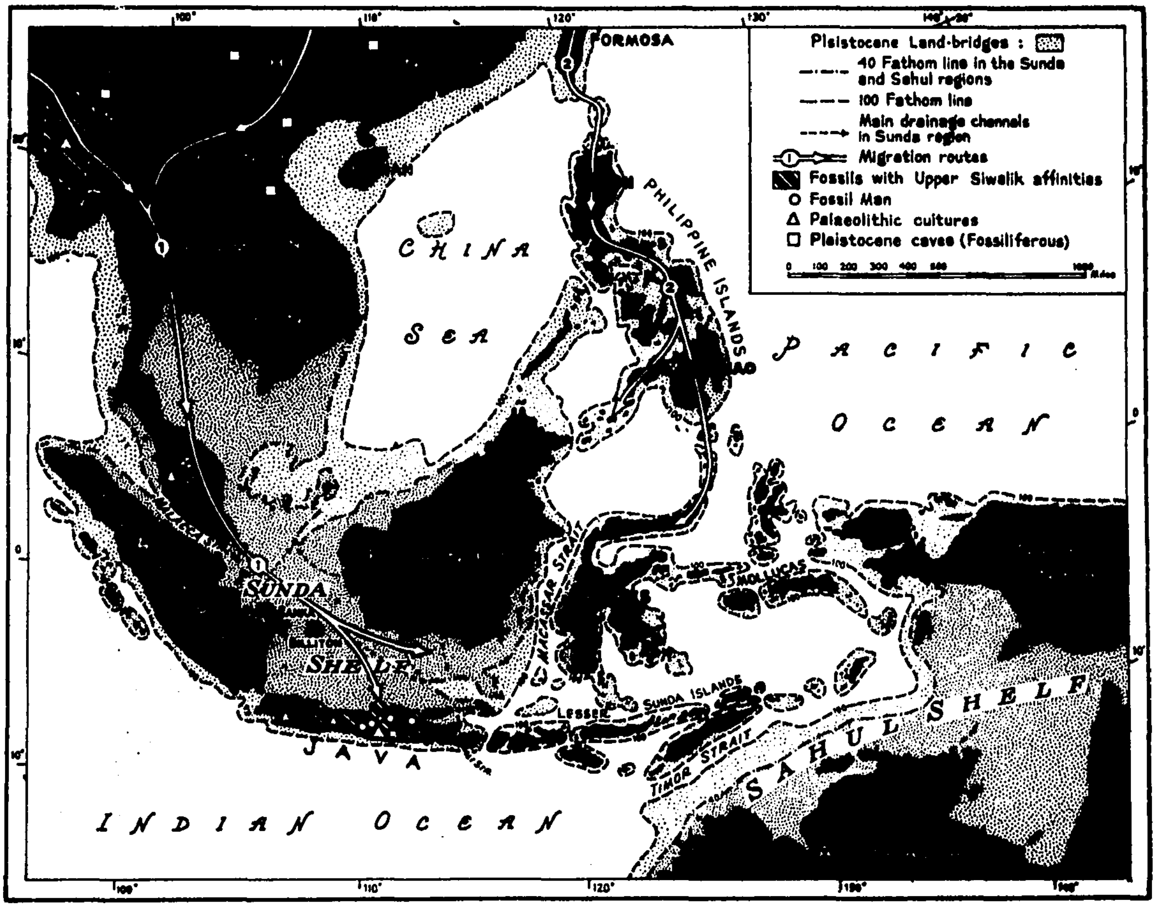


Figure 38. Pleistocene land-bridges connecting the Philippines and Asia (from Movius, 1949)

hypothesis noting that the fossil fauna of Taiwan contains virtually the same elements as that of the Philippines and that certain elements of the modern fauna are similar. Sartono (1973) describes the simultaneous Pleistocene uplift of parts of the Philippines and lowering of sea level and suggests that these factors resulted in a land-bridge between Taiwan and Northern Luzon. In addition to the faunal and geological evidence for a land-bridge with Taiwan, Fox and Peralta (1974) have noted that the mountains east, south, and west of the Cagayan Valley may have formed barriers to migration of large vertebrates from the south and that the vertebrates must have, therefore, migrated from the north. The effectiveness of the mountains as barriers to migration can be evaluated by considering the extent of the various coastal migration routes. These routes would have been very limited even during the lower sea levels of the Pleistocene. This is because there are almost no shallow shelves off the east and west coasts of Luzon as indicated by bathymetric data (Figure 39). Instead the sea floor drops to depths of 1,000 m or more within 10 km of the coast. Because of the limited shelf area for coastal migration routes, the mountains of Northern Luzon may have formed an effective barrier to migration as suggested by Fox and Peralta (1974).

The possible land-bridge between Luzon and Taiwan is formed by a submerged ridge, the North Luzon Ridge (Mammerickx et al., 1976), which now has a short, 2,000 m deep gap in it (Figure 39). Considering the active Pleistocene tectonics of the region, faults such as those mapped on the ridge (Figure 2) by Karig (1973) may easily have destroyed any Pleistocene land-bridge. Further studies of the newly recovered vertebrate fossils may provide the necessary evidence to conclusively locate the Pleistocene land-bridge or land-bridges.

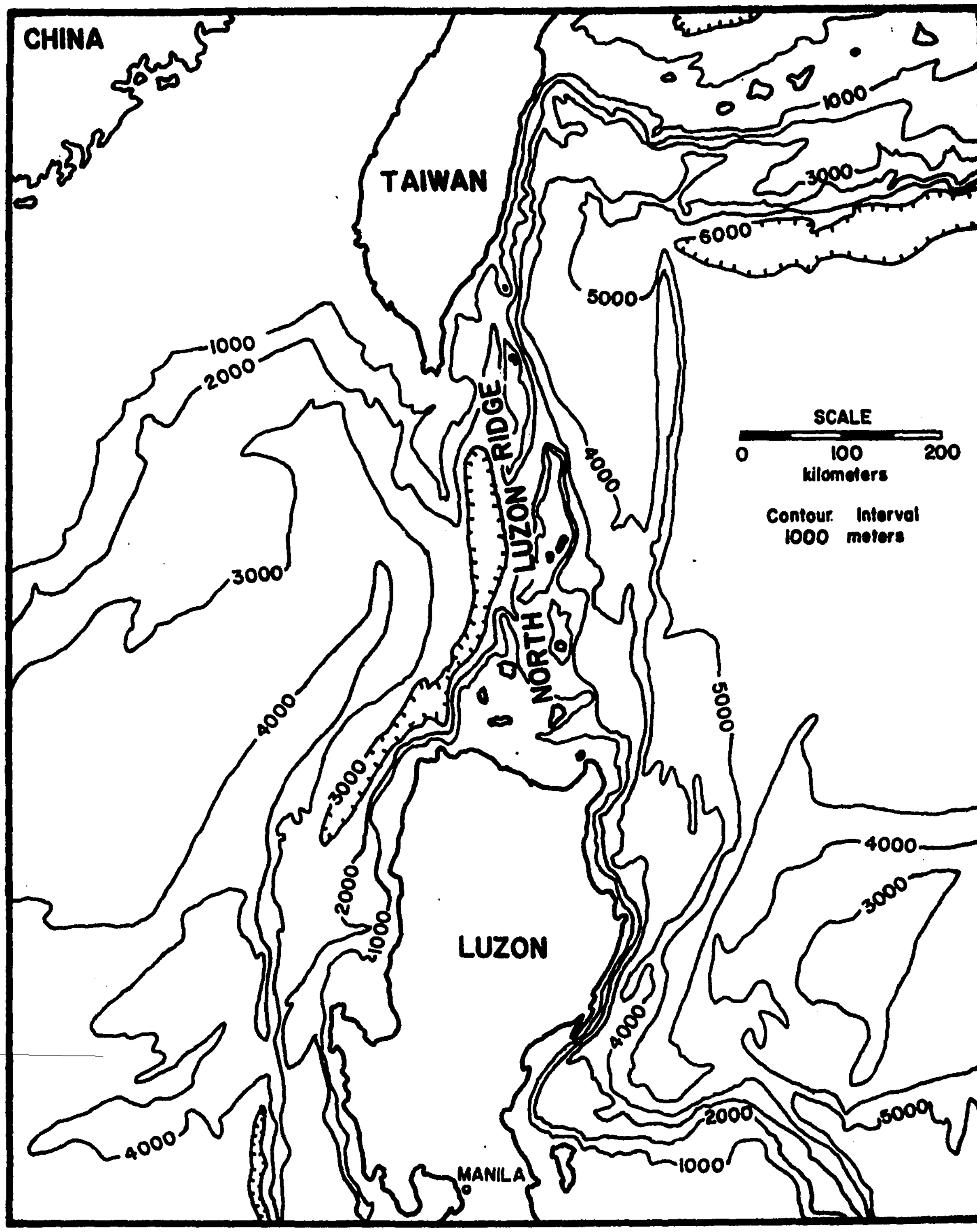


Figure 39. Bathymetric map of the Luzon-Taiwan region (after Mammerickx et al., 1976)

FACIES AND ENVIRONMENTS OF DEPOSITION

The Plio-Pleistocene sediments of the central Cagayan Valley are divided into five major lithofacies. The facies and respective depositional environments as inferred from lithologies, lithologic sequences, and sedimentary structures are 1) the interbedded fine grained sandstone and mudstone (SF) facies: delta front distal bar and distributary mouth bar; 2) the lenticular cross bedded, medium grained sandstone and siltstone (SS) facies: delta plain distributary channel, levee, and flood basin; 3) the polymictic conglomerate, trough cross bedded sandstone and claystone (GSF) facies: low energy fluvial channel and flood plain; 4) the clast-supported polymictic conglomerate and sandstone (GS) facies: high energy channel bar and gravel sheet; and 5) the massive matrix-supported pebble to boulder conglomerate, tuff-breccia and tuff (GT) facies: volcanic mudflows and debris flows (lahars), pyroclast flow (ignimbrite), and fall (tuff).

The vertical sequence and lateral distribution of facies record a regression of the Pliocene sea as the basin filled with detritus from the surrounding volcanic arcs. The first three facies mentioned above (SF, SS, GSF) occur in a vertical sequence throughout the central Cagayan Valley and document the transition from deltaic to fluvial sedimentation during the Pliocene. The last two facies overly and interfinger with the fluvial GSF facies along the western side of the Valley and represent an alluvial fan which formed at the base of the Cordillera Central as a result of Pleistocene tectonic and volcanic activity.

Lithofacies Codes

A lithofacies code system was introduced by Miall (1977, 1978) to standardize lithologic descriptions and facilitate comparisons between different fluvial sequences. The code consists of a capital letter designating the dominant grain size followed by one or two lower case letters which refer to the principal sedimentary structures. Only lithofacies of braided stream deposits were assigned codes by Miall (1978). The lithofacies codes of Miall (1978) are used in this report to refer to equivalent lithofacies of braided stream deposits in the Awidon Mesa Formation. In addition, code names are erected for lithofacies of the deltaic, meandering stream, and pyroclastic deposits of the Ilagan and Awidon Mesa Formations. The facies codes used in this report, including the codes defined by Miall (1978), are listed in Table 14. These codes refer to specific subfacies of the five major ~~lithofacies~~ defined in this chapter. The major lithofacies are also given code designations, capital letters which reflect the dominant grain size.

Interbedded Fine Grained Sandstone and Mudstone (SF) Facies

Description

This facies is the dominant facies of the Lower Member of the Ilagan Formation and is well-expected along the Pinacanauan de Tuguegarao River reaching a thickness of 120 meters where it overlies the prodelta claystones of the Baliwag Formation and forms a resistant escarpment along the river (Figure 40a). Slumps are common in most outcrops and may have occurred during deposition as well as after uplift and erosion. The facies is also exposed at Pangul Anticline where it overlies prodelta claystones

Table 14. Lithofacies codes and lithofacies of the Ilagan and Awidon Mesa Formations

Facies code	Lithofacies	Sedimentary structures	Interpretation
Gms[a]	Gravel, massive matrix-supported	None, normal or reverse grading	Lahars (including both mudflows and debris flows
Gm[a]	Gravel, massive or crudely bedded	Horizontal bedding, may have crude large scale trough cross bedding	Channel lag
Gt[a]	Gravel stratified	Large scale trough cross bedding	Minor channel fills
St[a]	Sand, medium to very coarse	Solitary (theta) or grouped (Pi) trough crossbeds	Dunes (lower flow regime)
Sp[a]	Sand, medium to very coarse	Solitary (alpha) or grouped (omikron) planar crossbeds	Transverse bars, sand waves (lower flow regime)
Sr[a]	Sand, very fine to medium grained	Ripple marks, small scale trough cross-beds	Ripples (lower flow regime)
Sh[a]	Silt, sand, very fine to very coarse, may be pebbly	Parallel laminations plane bedding	Planar bed flows (lower and upper flow regime)
Sl[a]	Sand, fine to medium grained	Low angle ($<10^{o}$) trough cross beds	Scour fills, antidunes
Sc	Sand, fine grained lenticular	Climbing ripple laminations, may appear structure-less	Crevasse splay
Scf	Sand, medium grained lenticular	Channel fill cross bedding	Distributary channel

[a]Lithofacies defined by Miall (1978).

Table 14. (continued)

Facies code	Lithofacies	Sedimentary structures	Interpretation
Sf	Sand, minor amounts of mud	Flaser bedding, bioturbated	Delta front distal bar
Spl	Sand, very fine to fine grained	Parallel laminations bioturbated	Delta front distal bar
SFi	Sand and mud, thinly interbedded	Parallel and wavy laminations, flaser and lenticular bedding, bioturbated	Delta front distal bar
Flb	Mud, sand	Lenticular bedding bioturbated	Delta front distal bar
Fl[a]	Sand, silt, mud	Laminated, very small ripples	Overbank or waning flood deposits
Fsc[a]	Silt, mud	Laminated to massive, bioturbated	Delta plain flood basin
Fm[a]	Mud, silt	Massive	Overbank or drape deposits
Fc	Clay, may be sandy	Massive	Meandering stream floodplain
Pp	Pisolitic claystone, iron pisolites	Pedogenic features; blocky structure	Paleosol
Tms	Tuff-breccia, massive matrix-supported	Normal grading of lithics, reverse grading of pumice gas escape structures	Pyroclast flow (ignimbrite)
Tt	Tuff, fine to coarse grained	Trough cross beds	Ground surge
Ts	Tuff, well sorted	Stratified	Pyroclast fall (airfall tuff)
Tr	Tuff, well sorted	Small scale trough cross beds	Reworked tuff

Figure 40. Photographs of the interbedded fine grained sandstone and mudstone (SF) facies and the lenticular crossbedded medium grained sandstone and siltstone (SS) facies

a. The SF facies exposed along the Pinacanauan de Tuguegarao River

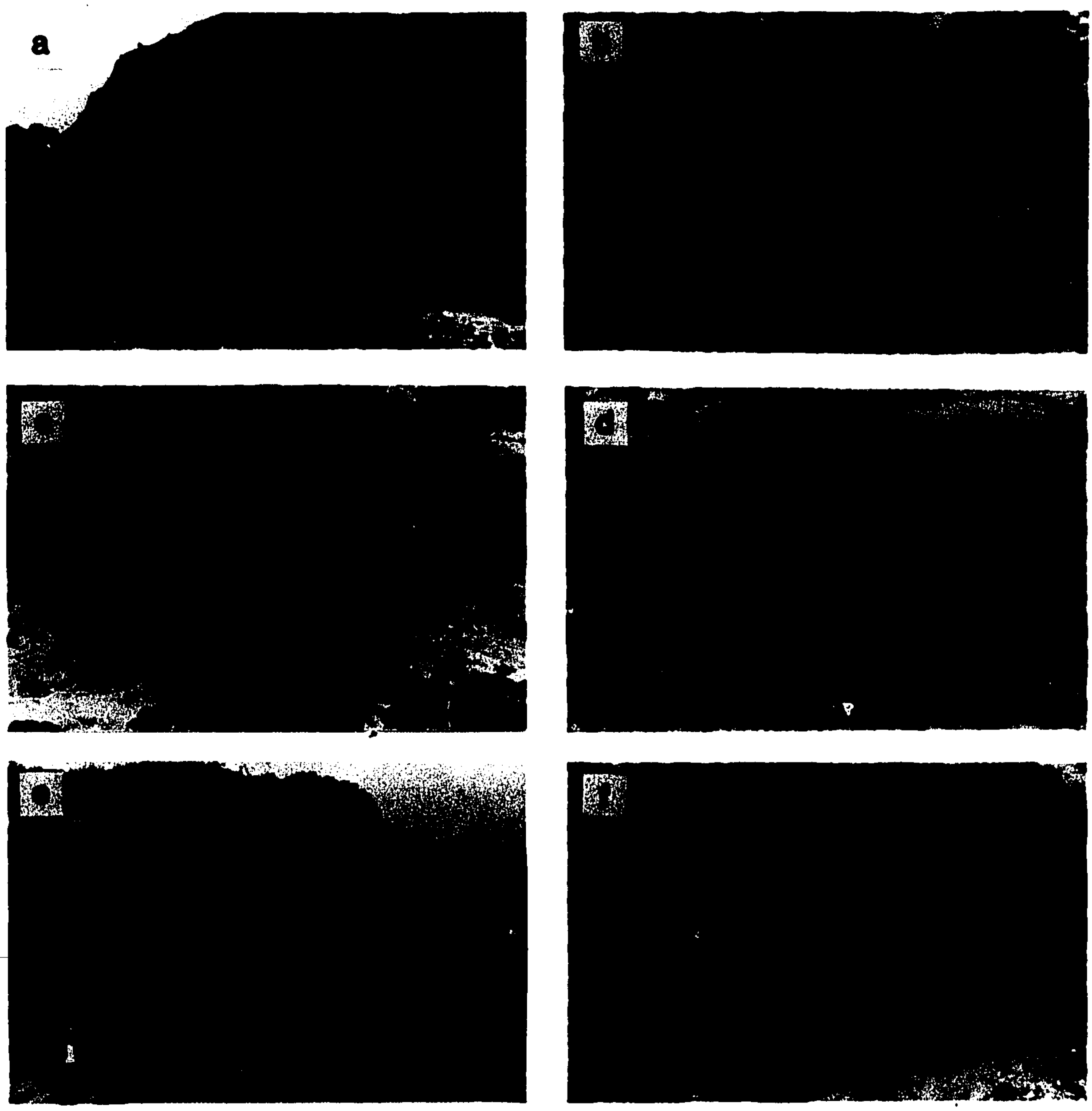
b. Lenticular and flaser bedding of the SFi subfacies of the SF facies

c. Bioturbation; vertical burrows typical of the SF facies

d. Gastropod and pelecypod fossils in sandstone of the Sf subfacies of the SF facies

e. Planar cross beds of the Sp subfacies of the SS facies

f. Lenticular channel fill cross bedded sandstone Scf subfacies of the SS facies


a

of the Mabaca River Group. In the upper part of the Lower Member of the Ilagan Formation, the facies interfingers with and is overlain by the SS facies. The SF facies is composed of two main subfacies and three minor subfacies as illustrated in Figure 41.

The main subfacies of the SF facies are the thinly interbedded sandstone and mudstone (SFi) and the fine to medium grained trough cross bedded sandstone (St) subfacies. The SFi subfacies is the dominant subfacies and is composed of very thin to thin, interbedded dark greenish gray (5GY4/1) to yellowish orange (10YR8-6) well-sorted sandstones and mudstones with subordinate siltstones. The sandstones and mudstones which usually exhibit parallel or wavy laminations, or flaser and lenticular bedding (Figure 40b), are of constant thickness and laterally continuous for at least several meters. The SFi subfacies is commonly several meters thick. The St subfacies is composed of yellowish orange (10YR8/6) fine to medium grained sandstones which commonly appear structureless. The sandstones, which sometimes display faint small scale trough cross bedding and occasional flaser bedding, are of uniform thickness and form laterally continuous sheet sands. Most sandstones are .5 to 2 m thick, but several thicker beds occur with a maximum thickness of 10 m.

The minor subfacies of the SF facies are the fine grained parallel laminated sandstone (Spl), the flaser bedded sandstone (Sf), and the lenticular bedded mudstone (Flb) facies. These facies are composed of yellowish orange (5GY4/1) sandstones and very pale orange (10YR8/2) mudstones which are usually less than 1 m thick but may be up to 3 m thick. The Flb facies contains both single and connected lenses up to 2 cm thick.

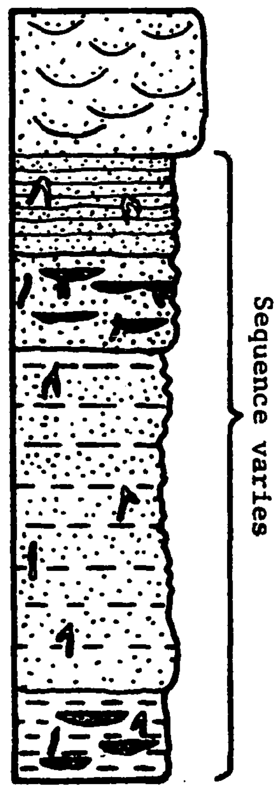
SUBFACIES	FACIES CODE	VERTICAL SEQUENCE	SEDIMENTARY STRUCTURES	THICKNESS	INTERPRETATION
Sand, fine to medium grained	St		Trough cross bedded, commonly appears structureless	.5-10 m	Distributary mouth bar
Sand, very fine to fine grained	Spl		Parallel laminated bioturbated	4 m	Distal bar
Sand, mud	Sf		Flaser bedding, bioturbated	6 m	Distal bar
Sand and mud, thinly interbedded	SFi		Parallel and wavy laminations, flaser and lenticular bedding, bioturbated	11 m	Distal bar
Mud, sand	Flb		Lenticular bedding, bioturbated	3 m	Distal bar

Figure 41. Subfacies of the interbedded fine grained sandstone and mudstone (SF) facies (the main subfacies are underlined)

Flasers of the SF facies are also either single or connected but usually not more than 1 cm thick.

Subfacies of the SF facies occur in vertical sequences that coarsen upward as illustrated in Figure 41. The SFi subfacies is the dominant subfacies of the lower and middle part of the sequence and is interbedded with the Flb, Sf, and Spl subfacies. Sandstones of the Sf and Spl subfacies become more abundant and thicker in the upper part of the sequence which is capped by sandstones of the St subfacies.

A variety of fossils occur in the SF facies. Plant fragments are abundant along many bedding planes while burrows up to 2.5 cm in diameter (Figure 40c) are common throughout the facies. The sandstones commonly contain forams and minor amounts of glauconite. Fractured or abraded pelecypods and gastropods also occur and are concentrated in sandstones of the St and Sf subfacies along with shell molds. They may be concentrated along bedding planes or may be distributed throughout the sandstones in varying orientations as in Figure 40d. Shark's teeth also occur in the facies but are rare. Calcareous and ferruginous, disk to blade shaped concretions up to boulder size are often concentrated along horizons that are fossiliferous suggesting the calcite was derived from dissolution of carbonate-rich fossils.

Depositional environment

The SF facies is interpreted as delta front distal bar and distributary mouth bar deposits based on the lithology, sedimentary structures, thickness, lateral continuity, vertical sequence and fossil content of the subfacies, and on facies association. The sandstones and mudstones of the

SFi, Sf, Spl, and Flb subfacies are distal bar deposits. Interbedded, bioturbated sandstones and mudstones with parallel and wavy laminations and flaser and lenticular bedding are characteristic of delta front distal bars as described by Coleman and Gagliano (1965), Donaldson, Martin, and Kanes (1970), and Miall (1979). The thicker sandstones of the St subfacies which exhibit small scale trough cross bedding and often contain fractured and abraded fossil molluscs are interpreted as distributary mouth bar deposits as described by Coleman and Gagliano (1965). The lateral continuity of the sandstones of the St and Sf subfacies suggests that they may be sheet sands (Reineck and Singh, 1975) which formed by seaward progradation of the delta. Coarsening upward sequences in the SF facies, which are from 30 to 80 m thick, also reflected progradation of the delta. Miall (1979) notes that coarsening upward cycles of delta lobes (typically 50 to 100 m thick) are one of the most characteristic features of deltaic sediments. The interpretation of the SF facies as delta front deposits is consistent with the associated facies. Delta front deposits characteristically overlie prodelta deposits such as those of the underlying Baliwag Formation and underlie fluvial sediments such as those of the Ilagan Formation.

The characteristics of delta front deposits are used to define delta types (Elliot, 1978). Deposits of the SF facies probably reflect development of a high constructive lobate delta (Figure 42) as defined by Fisher et al. (1969). High constructive deltas are river dominated deltas which are characterized by thick delta front deposits. These deposits accumulate as the delta advances by the continuous addition of sediment to the delta front by distributary channels (Miall, 1979). Slumping commonly occurs in these deposits from oversteepening of the delta front slope (Coleman et al.,

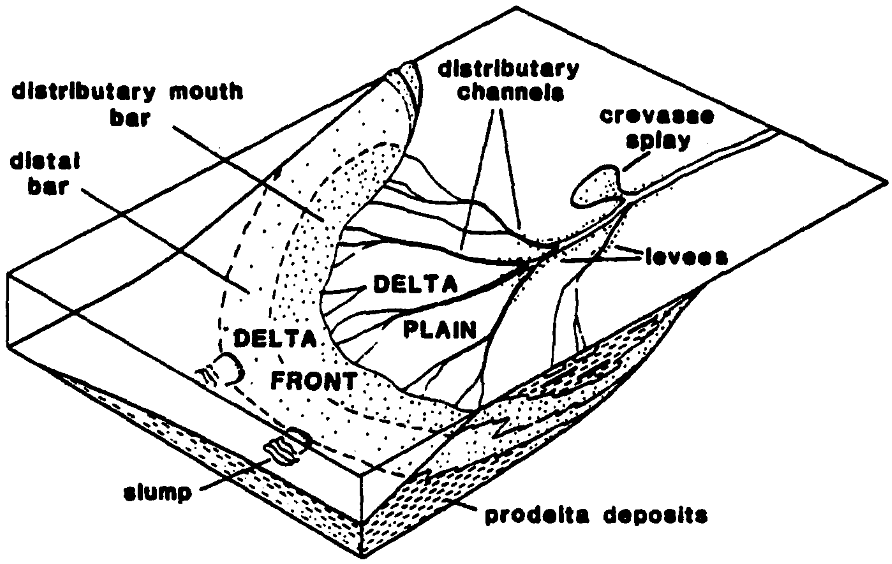


Figure 42. Block diagram of a lobate high constructive delta

1974). Lobate high constructive deltas are characterized by sheet sands which reflect low subsidence rates and lack bar finger sands characteristic of elongate (bird foot) deltas which have higher subsidence rates (Fisher et al., 1969). The Cagayan delta front deposits of the SF facies are interpreted as part of a high constructive lobate delta system based on the thickness of the facies, the coarsening upward sequence, and the abundance of sheet sands in contrast to bar-finger sands. The low subsidence rates reflected by the lobate delta system are consistent with the interpretation that the Cagayan basin was being uplifted during the Plio-Pleistocene.

Lenticular Cross Bedded Medium Grained Sandstone and Siltstone (SS) Facies

Description

The SS facies occurs in the upper part of the Lower Member of the Ilagan Formation where it interfingers with and overlies the SF facies. The only exposures occur at Pangul Anticline and along the upper part of the escarpment along the Pinacanauan de Tuguegaro River where the maximum thickness measured is 24 meters. The preserved thickness of this facies may be greater at Pangul Anticline, but abundant slumping and vegetative cover preclude accurate measurement.

The SS facies is composed of yellowish gray (5Y7/2) to dark yellowish orange (10YR6/6) sandstones, siltstones, and subordinate mudstones which are divided into 5 subfacies as indicated in Figure 43. Massive, lenticular, medium to coarse grained sandstones up to 9 m thick form the Scf subfacies. The sandstones display small or large scale trough cross bed sets up to 50 cm thick, or channel fill cross bedding (Reineck and Singh, 1975), and have scoured bases which cut into finer grained deposits (Figure 40e). The Sp subfacies is formed by planar cross bedded medium grained sandstones that occur in sets up to 30 cm thick (Figure 40f) and is a minor subfacies of the SS facies. The Sc subfacies is also a minor subfacies and is composed of lenticular structureless appearing sands up to 40 cm thick which are probably ripple bedded. The Scf, Sp, and Sc subfacies are interbedded with finer grained sediment of the Fsc and Sfi subfacies. The Fsc subfacies is composed of laminated to massive siltstones and mudstones. Interbedded parallel or ripple laminated sandstones and mudstones form the SFi subfacies which is commonly associated with the sandstones of the Scf and Sp subfacies in the vertical succession of subfacies. The vertical sequence of subfacies as illustrated in Figure 43 is not as regular as the vertical sequence of the SF facies.

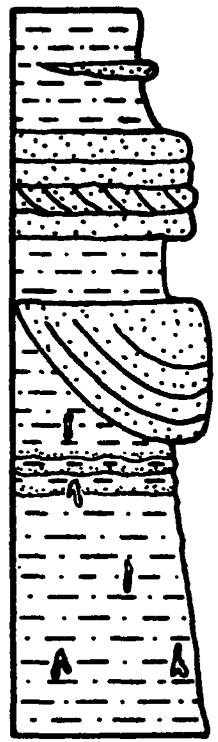
SUBFACIES	FACIES CODE	VERTICAL SEQUENCE	SEDIMENTARY STRUCTURES	THICKNESS	INTERPRETATION
Sand, fine grained lenticular	Sc		Appears structureless	40 cm	Crevasse splay
Sand, medium grained	Sp		Planar cross beds (lower flow regime)	30 cm	Distributary channel
Sand, medium grained lenticular	Scf		Trough cross beds, channel fill cross bedding	9 m	Distributary channel
Sand and mud, interbedded	SFi		Parallel or ripple laminated sand bioturbated	2 m	Levee
Silt, mud	Fsc		Laminated to massive bioturbated	11 m	Floodbasin

Figure 43. Subfacies of the lenticular cross bedded medium grained sandstone and siltstone (SS) facies

Depositional environment

The SS facies is interpreted to represent delta plain distributary channel, levee, and floodbasin deposits (Figure 42). Subaqueous distributary channels and levees were included in this facies but were not differentiated. Sandstones of the Scf and Sp subfacies are interpreted as distributary channel sandstones. Lenticular cross bedded distributary channel sandstones with scoured basal surfaces have been described by many authors and are summarized by Coleman and Gagliano (1965) and Wright (1978). The finer grained sediments of the SFi and Sc subfacies represent levee and crevasse splay deposits. Fine grained laminated sands are commonly deposited along distributary channel margins as levees while lenticular sandstones are deposited by crevasse splays from crevasse channels in the levee crests (Elliot, 1978). The laminated to massive bioturbated silts and muds of the Fsc subfacies are flood basin deposits which include marsh, swamp, interdistributary bay, and mud flat deposits (Coleman and Gagliano, 1965). Flood basin deposits typically contain large amounts of organic matter which may form peat or coal. While no peat or coal deposits were found in outcrops of the facies examined during this study, they have been reported to occur in the Ilagan Formation by Corby et al. (1951). In addition to the evidence from the various subfacies, the stratigraphic occurrence of the facies supports the delta plain environmental interpretation. The facies interfingers with and overlies delta front deposits and interfingers with fluvial conglomerates and sandstones of the GSF facies.

Polymictic Conglomerate, Trough Cross Bedded Sandstone and Claystone (GSF) Facies

Description

This facies is the most extensive of the Plio-Pleistocene deposits. It is exposed along Enrile and Pangul anticlines where it forms the Upper Member of the Ilagan Formation and at the latter, reaches a thickness of 500 meters (Figure 44a). This facies also occurs in the Awidon Mesa Formation where it is interbedded with the polymictic clast-supported conglomerate and sandstone (GS) facies and the massive matrix-supported pebble to boulder conglomerate, tuff-breccia, and tuff (GT) facies. In the northern part of Cabalwan Anticline, it reaches a thickness of 120 meters.

The GSF facies is composed of pale yellowish orange (10Y8/6) granule to cobble conglomerates, coarse to fine grained sandstones and siltstones, and pale olive (10Y6/2) claystones which characteristically occur in fining upward sequences 5 to 40 m thick. The facies is divided into 7 subfacies as illustrated in Figure 45. Massive or large scale trough cross bedded granule to cobble polymictic conglomerates form the Gm subfacies which occurs at the base of most sequences. The conglomerates have sharp but irregular scoured basal contacts and sometimes display trough cross bed sets up to 1 m thick (Figure 44b). The older Pliocene conglomerates of this subfacies are thin channel lag conglomerates composed primarily of granules and pebbles at the base of thick sandstones. The Plio-Pleistocene conglomerates and conglomeratic sandstones are thicker, up to 7 meters, and contain coarser pebble to cobble size clasts. These are dominantly disk to equant in shape, subrounded to rounded, and composed of basalt and porphyritic andesite with minor amounts of quartzite, chert, and sedimentary rock

Figure 44. Photographs of the polymictic conglomerate, trough cross bedded sandstone, and claystone (GSF) facies

a. The GSF facies as exposed at Pangul Anticline

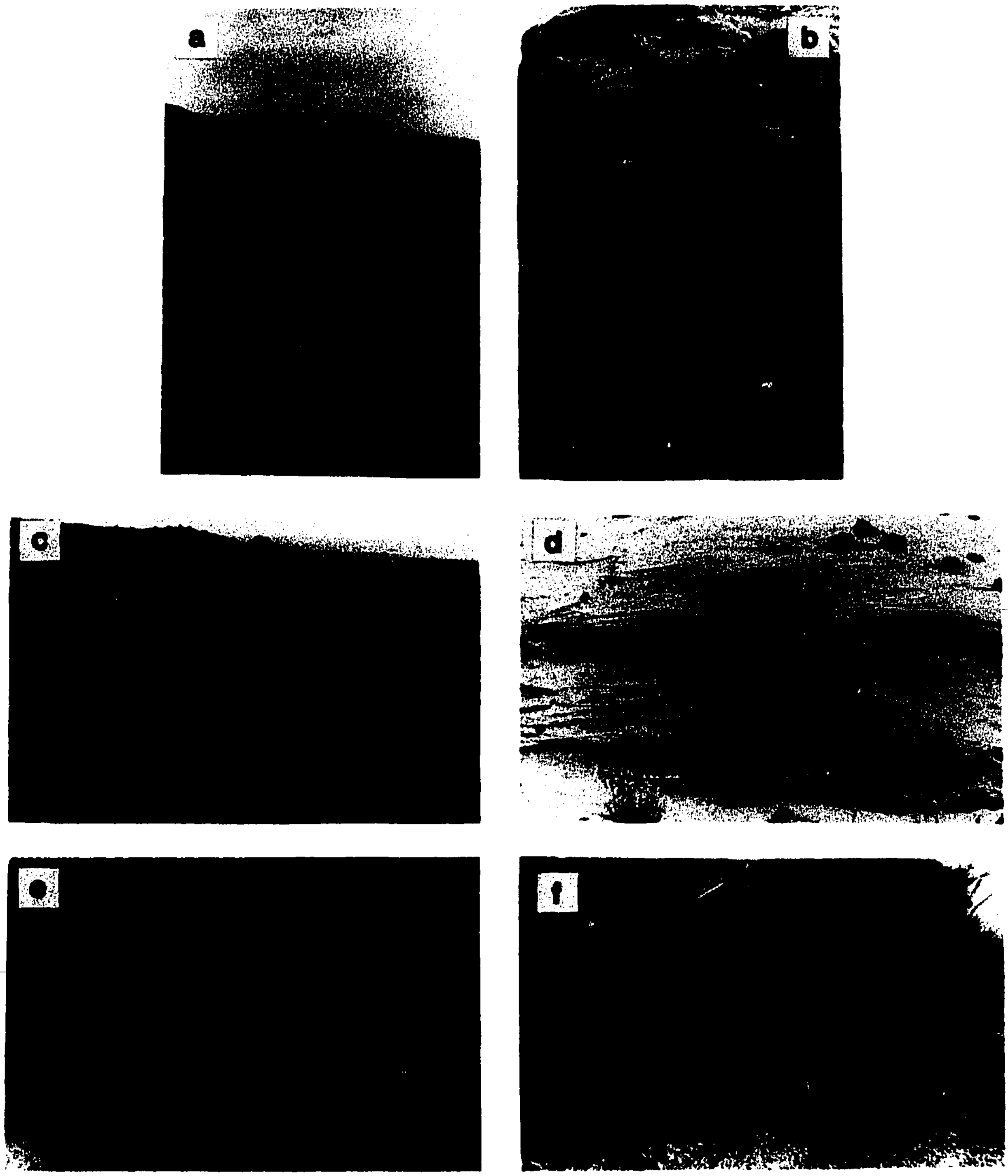
b. Polymictic pebble conglomerate, the Gm subfacies, overlain by trough cross bedded sandstone of the St subfacies at Pangul Anticline

c. 8 m thick trough cross bedded sandstone of the St subfacies at Pangul Anticline

d. Large scale trough cross bedding, St subfacies; Pangul Anticline

e. Small scale trough cross bedding and climbing ripple laminations of the Sc subfacies at Cabalwan Anticline

f. Iron-rich pisolite paleosol of the Pp subfacies, in upper part of a Fc subfacies claystone at Pangul Anticline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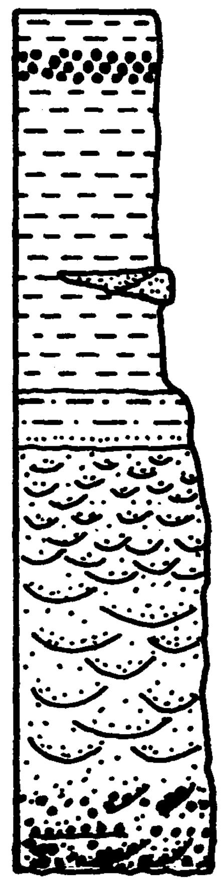
SUBFACIES	FACIES CODE	VERTICAL SEQUENCE	SEDIMENTARY STRUCTURES	THICKNESS	INTERPRETATION
Pisolitic claystone, iron pisolites	Pp		Pedogenic features; subangular blocky structure	2 m	Paleosol
Clay, mud, may be sandy	<u>Fc</u>		Massive	2-35 m	Floodplain
Sandy, fine grained lenticular	Sc		Climbing ripple laminations, may appear structureless	1 m	Crevasse splay deposit
Silt, very fine sand	Sh		Parallel laminations	2 m	Planar bed flow l. and u. flow regime
Sand, very fine to medium grained	<u>Sr</u>		Small scale grouped (pi) trough cross beds	1-3 m	Ripples (lower flow regime
Sand, medium to very coarse	<u>St</u>		Large scale grouped (pi) trough cross beds	2-8 m	Dunes (lower flow regime)
Gravel, massive or crudely bedded	<u>Gm</u>		Horizontal bedding may have crude large scale trough cross bedding, scoured base	1-7 m	Channel lag

Figure 45. Subfacies of the polymictic conglomerate, trough cross bedded sandstone, and claystone (GSF) facies (main subfacies are underlined)

fragments. The conglomerates of the Gm subfacies are overlain by trough cross bedded sandstones of the St subfacies (Figure 44b). The sandstones, which are up to 8 m thick (Figure 44c), fine upward from very coarse to medium sand and exhibit large scale trough cross bed sets up to 1 m thick (Figure 44d). The Sf subfacies commonly grades to finer grained sandstones of the Sr subfacies that display small scale trough cross bedding. The Sr subfacies is commonly only 1 to 3 m thick. Valid paleocurrent estimates were not obtainable from the St and Sr subfacies due to limited exposures and the highly variable directional nature of the trough cross bedding. Parallel or wavy laminated very fine grained sandstones and siltstones of the Sh subfacies characteristically overlie the Sr subfacies. These deposits are usually thin, rarely exceeding 2 m. Occasional thin parallel laminated fine grained sands also occur within the Sr subfacies at various levels. The fining upward sequences are capped by massive claystones of the Fc subfacies. This subfacies ranges from 2 to 35 m in thickness with an average thickness of 20 m. The claystones, which are often sandy or contain thin sand lenses, are usually poorly exposed to covered as in Figure 44a. Fine grained lenticular sandstones, the Sc subfacies, commonly occur in the claystone sequences. These sandstones often appear structureless but usually display faint ripply bedding or climbing ripple laminations (Figure 44e). Numerous claystones also contain horizons with blocky structure or iron rich pisolites and are designated as the Pp subfacies. These horizons which are up to 2 m thick are laterally continuous along strike of the beds as seen in Figure 44f.

The conglomerates and sandstones of the GSF facies form resistant, laterally extensive tabular deposits of nearly uniform thickness. They are commonly traceable in scarp faces of hogbacks along the flank of anticlines for at least several kilometers (Figure 44a).

Plant and vertebrate fossils are locally abundant in the GSF facies. Permineralized wood is commonly found in tuffaceous sandstones, siltstones, and claystones of the GSF facies in the Ilagan Formation. The Awidon Mesa Formation, in contrast, lacks permineralized wood in the GSF facies but contains in situ vertebrate fossils in conglomerates, sandstones, and claystones.

Depositional environment

The GSF facies is interpreted to have been deposited by a meandering stream system, as illustrated in Figure 46. The conglomerates of the Gm subfacies and sandstones of the St, Sr, and Sh subfacies are lateral accretion deposits which accumulated as point bars of intermediate to high sinuosity streams. The major characteristics of lateral accretion or point bar deposits have been well-documented by numerous investigations of modern and ancient sediments (Allen, 1965, 1970; Visher, 1965; McGowan and

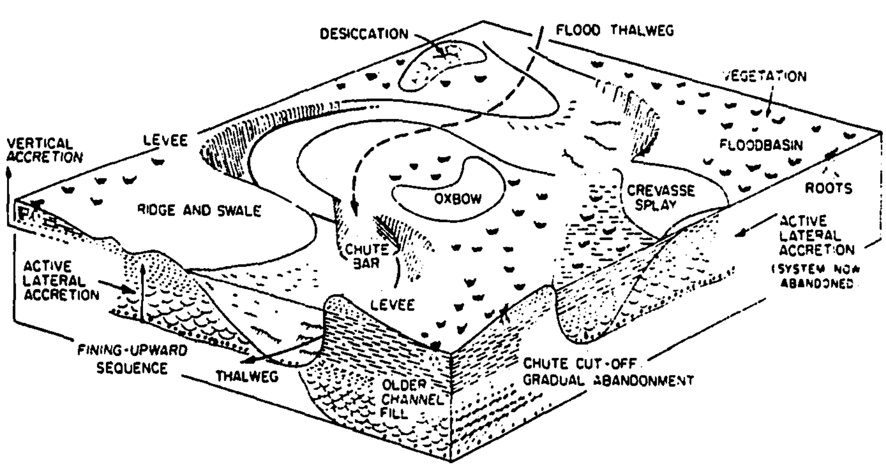


Figure 46. Block diagram of a meandering stream system (from Walker and Cant, 1979)

Garner, 1970; Walker and Cant, 1979). The conglomerates of the Gm subfacies represent channel lag in the bottom of migrating channels and were overlain by trough cross bedded sand of the Sf and Sr subfacies as the river migrated laterally by eroding the concave bank. As described by Harms and Fahnstock (1965), the flow phenomena range from the upper part of the lower flow regime to the lower part of the lower flow regime. The parallel laminated fine grained sandstones and siltstones of the Sh subfacies which overlie or occasionally occur in the Sr subfacies indicate flow over a plane bed at high velocities in the upper flow regime of Harms and Fahnstock (1965). The massive claystones of the Fc subfacies, which are often sandy, represent vertical accretion deposits which formed on floodplains outside the main channel. The clays and minor amounts of silt and sand settled from suspension thus building up the floodplain. Lenticular fine grained sandstones of the Sc subfacies which occasionally occur in floodplain claystones are interpreted as crevasse splay deposits. Leeder (1974) has suggested that lenticular sandstones in floodplain deposits may be interpreted as crevasse splay deposits particularly when laterally limited in extent. The climbing ripple laminations of some deposits are also common primary structures in crevasse splay deposits (Collinson, 1978a). Horizons in the Fc subfacies with iron rich pisolites and/or blocky structure, the Pp subfacies, are interpreted as paleosols. These horizons much have formed in stable, upland interchannel areas which were subaerially exposed long enough for iron enrichment of the soil and the development of blocky structure.

The geometry of the sandstones of the St and Sr subfacies and claystones of the Fc subfacies provides further evidence regarding the nature of the stream system. The sandstones are laterally continuous, hundreds of meters to several kilometers in length, and fairly uniform in thickness. Modern meandering streams deposit similar sheet-like sand bodies as they migrate across floodplains (McGowan and Garner, 1970). Moody-Stuart (1966) has suggested that sheet-like sandstone geometry is a diagnostic characteristic of meandering stream channel sands. The sandstones are thicker, 6-12 m, in the Upper Member of the Ilagan Formation and thinner, 2-6 m, in the Awidon Mesa Formation. The Ilagan Formation sandstones were deposited by a large meandering stream with a channel up to 12 m deep. This stream flowed to the north along the axis of the valley. The thinner Awidon Mesa Formation sandstones represent smaller meandering tributary streams which flowed along the western side of the valley. Ratios of thickness between lateral accretion and vertical accretion deposits have been used to try to interpret characteristics of the stream systems during deposition. In this facies, the ratios range between 1:2 and 1:3 as the vertical accretion deposits are 2 to 3 times thicker than the lateral accretion deposits. Instead of providing further evidence on the nature of the stream system, these ratios most probably reflect the width of the alluvial plain and rates of subsidence and orogenesis as noted by Collinson (1978b) and Friend (1978).

Recent research on meandering stream deposits has indicated that they may be more complex than previously thought. Jackson (1978) notes that widely used criteria for meandering streams do not occur in all meandering streams or they can exist in nonmeandering streams. Epsilon cross-stratification, for example, which Moody-Stuart (1966) considers a diagnostic characteristic of laterally accreted meandering stream deposits, has not been observed in the Cagayan Valley. Observations by Allen (1970) and Jackson (1978) indicate that epsilon cross-stratification is rare in many modern and ancient deposits.

In summary, the interpretation of the GSF facies is based on the fining upward grain size, sedimentary structure sequences, geometry of the sand bodies, and thick flood plain deposits. The stratigraphic position of the facies between deltaic and braided stream deposits also supports the interpretation that the GSF facies represents meandering stream deposits.

Clast-Supported Polymictic Conglomerate and Sandstone (GS) Facies

Description

Conglomerates and sandstones of the GS facies occur throughout the Awidon Mesa Formation but are thicker, up to 50 m thick, and more numerous toward the southern and western part of the central Cagayan Valley. Finer grained mudrocks occur in some outcrops but form only a minor component of the facies. The GS facies is well-exposed along the flanks of the anticlines (Figure 47) where it overlies and interfingers with the GSF facies and interfingers with the GT facies.

The yellowish gray (5Y7/2) to moderate yellowish brown (10YR5/4) conglomerates and sandstones have been divided into 5 subfacies (Figure 48) which are equivalent to various alluvial gravel facies defined by Miall (1978). The Gm subfacies, massive clast supported pebble to boulder conglomerate, is the dominant subfacies (Figure 47b). The conglomerates are composed primarily of andesite and basalt clasts and display crude horizontal bedding and faint imbrication indicating flow from the west. The Gm subfacies increases in thickness from north to south from a maximum

Figure 47. Photographs of the clast-supported polymictic conglomerate and sandstone (GS) facies

a. The GS facies along road north of Minabbag, Kalinga-Apayao Province

b. Clast supported polymictic cobble conglomerate, the Gm subfacies, southern Enrile Anticline

c. Trough cross bedded polymictic pebble conglomerate of the Gt subfacies, Enrile Anticline

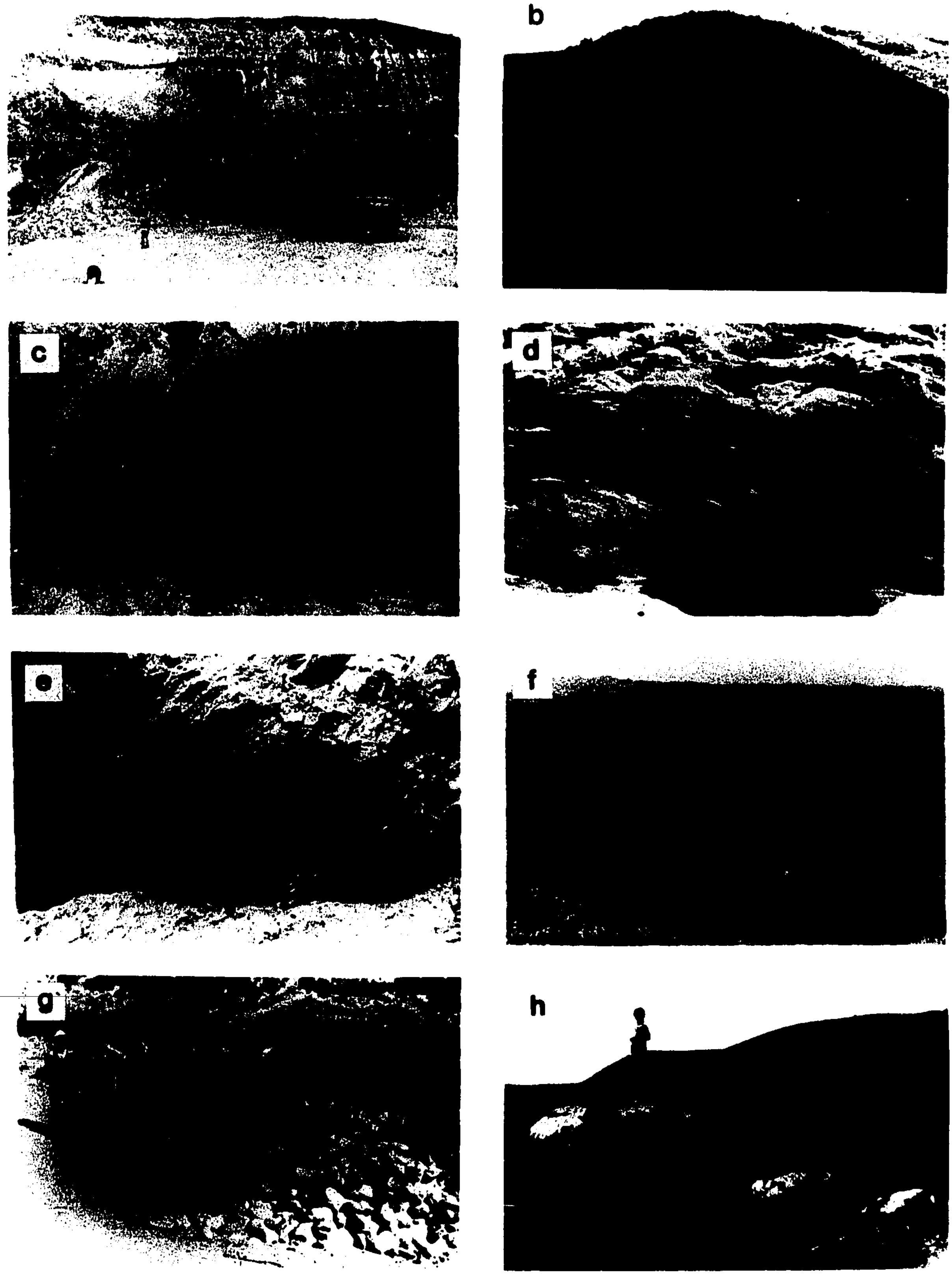
d. Planar cross bedding of the Sp subfacies, northern Enrile Anticline

e. Plane bedding of the Sh subfacies, southern Enrile Anticline

f. Tektite bearing lag gravel derived from pebble conglomerate of the Gm subfacies, west flank of Cabalwan Anticline

g. Boulders from lag gravels at Wanawan Ranch, east flank of Pangul Anticline

h. Very large boulders of southern Pangul Anticline. The largest boulder has a long diameter of 5.5 m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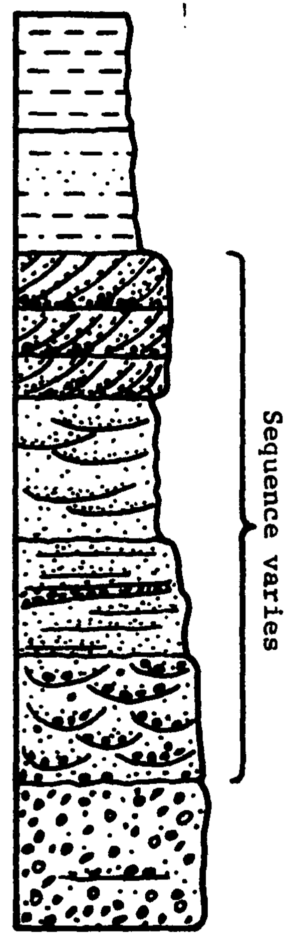


SUBFACIES	FACIES CODE	VERTICAL SEQUENCE	SEDIMENTARY STRUCTURES	THICKNESS	INTERPRETATION
Mud, silt	Fm		Massive	2 m	Overbank or drape deposits
Sand, silt, mud	Fl		Lamination, very small ripples	2 m	Overbank or waning flood deposits
Sand, medium to very coarse, conglomeritic	Sp		Solitary (alpha) or grouped (omikron) trough cross beds	1 m	Lingoid, transverse bars, sand waves (lower flow regime)
Sand, fine to medium	Sl		Low angle ($<10^{\circ}$) trough cross beds	6 m	Scour fills, antidunes
Sand, medium to very coarse conglomeratic	Sh		Plane bedding	6 m	Planar bed flow (upper flow regime)
Gravel, stratified	Gt		Large scale trough cross bedding	6 m	Minor channel fills
Gravel, massive pebble, cobble to boulder	Gm		Crude horizontal bedding and imbrication	22 m	Longitudinal bars lag deposits

Figure 48. Subfacies of the clast supported polymictic conglomerate and sandstone (GS) facies

of 6 m in the northern part of the study area to a maximum exposed thickness of 22 m in the south. The clast size also increases from north to south. Pebble conglomerates predominate in the north while cobble to boulder conglomerates are more common in the south. Trough cross bedded pebble conglomerates form the Gt subfacies (Figure 47c) which occurs throughout the northern and eastern part of the study area. The conglomerates grade upward to sandstone in each cross bed set. Convolute bedding is well-developed in some of the sandstones in this subfacies. Three relatively minor subfacies are formed by planar cross bedded conglomeratic sand, the Sp subfacies (Figure 47d), plane bedded sandstones and conglomerates of the Sh subfacies (Figure 47e), and low angle trough cross bedded sand, the Sl subfacies. The Sl and Sh subfacies are up to 6 m thick while the Sp subfacies attains a thickness of only 1 m with planar cross bed sets up to 30 cm thick. The subfacies just described occur in a vertically unordered sequence characterized by abrupt textural changes.

The increase in clast size of the Gm subfacies from north to south is best observed in lag gravels derived from the subfacies. In the north, lag gravels formed by pebble size clasts occur along the flanks of Enrile and Cabalwan (Figure 47f) anticlines. Farther to the south cobble to boulder size clasts are common along the northeast flank of Pangul Anticline. At the southern margin of the study area, medium to very large boulders occur (Figure 47h). Some of the larger boulders are probably derived from the Gms subfacies of the Gt facies which interfingers with the GS facies.

The conglomerates and sandstones of the GS facies are often interbedded with fine grained deposits in the eastern and northern parts of the study area. Mudstones with sand and silt laminations or small ripples form the Fl subfacies while more massive mudstones are classified as the Fm subfacies. These subfacies are usually less than 2 m thick. Some poorly exposed to covered horizons up to 20 m thick associated with the coarser subfacies of the GS facies appear to be formed by these subfacies.

Disarticulated vertebrate fossils occur in situ in the Gm, Gt, and Fl subfacies at Cabalwan Anticline. Numerous vertebrate fossils and tektites have also been found in the lag gravels derived from the Gm and Gt subfacies at northern Enrile, Cabalwan, and northern Pangul anticlines.

Depositional environment

The GS facies is interpreted as deposits of east flowing braided streams which formed part of an extensive Pleistocene alluvial fan complex at the base of the active Cardillera Central. The imbricated, horizontally bedded to massive, clast-supported conglomerates of the Gm subfacies were deposited primarily as proximal gravel sheets and low relief longitudinal or diagonal bars (Smith, 1974; Hein and Walker, 1977; Rust, 1978) on the upper to mid-fan (Figure 49). The large scale trough cross bedded conglomerates and sandstones of the Gt subfacies are interpreted to be more distal channel bar deposits (Figure 49). Rust (1979) has described similar large scale trough cross bedded conglomerates which fine upward as a response to channel migration or shallowing of the water over bars and channels as they were accreted in the distal reaches of braided streams. The planar cross bedded sands (Sp subfacies), plane bedded sands (Sh subfacies), and low angle trough cross bedded sands (Sl subfacies) are also interpreted as distal braided stream deposits. These deposits reflect a gradual decrease in the particle size and water depth ratio (Rust, 1978) as the stream

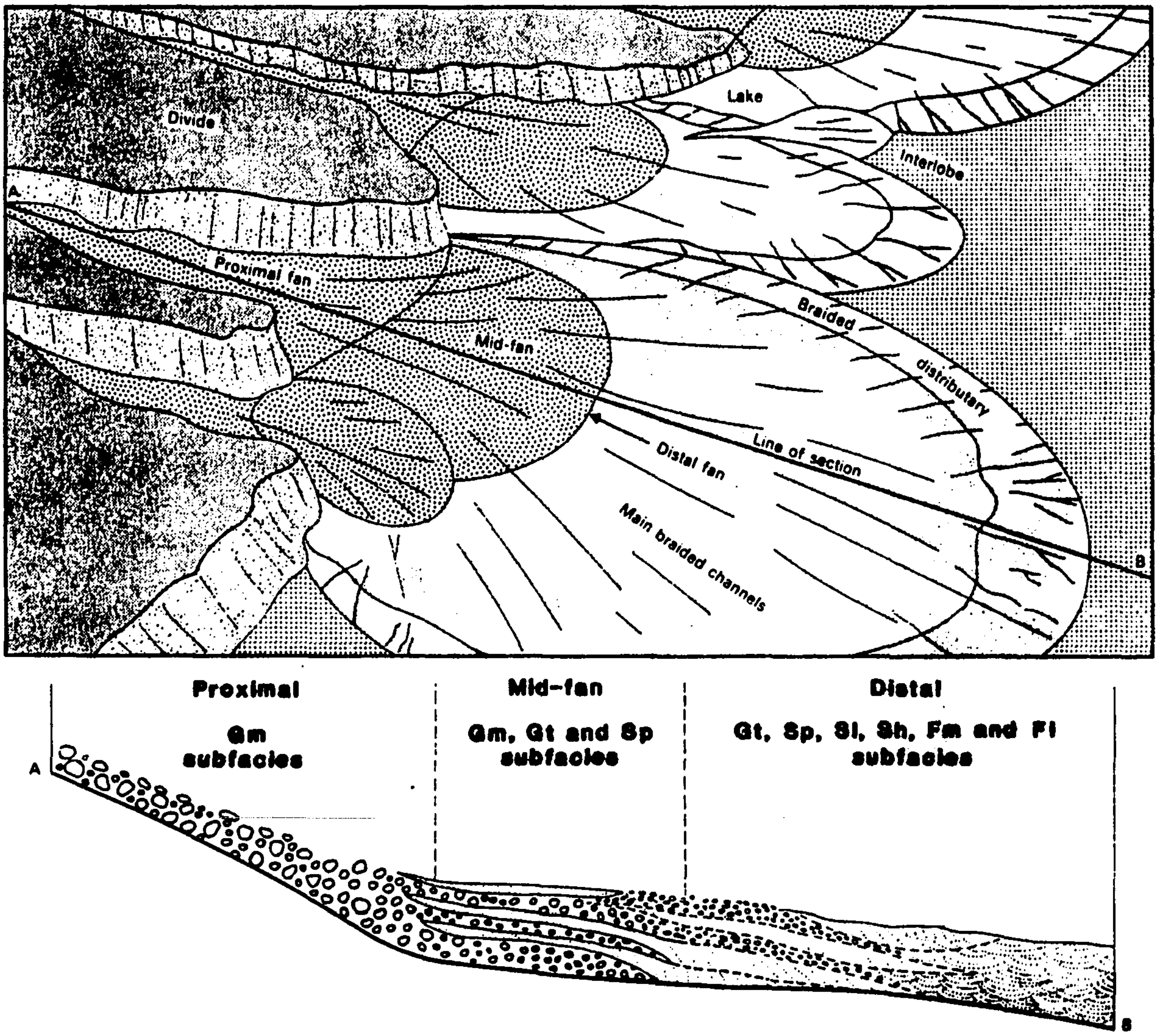


Figure 49. Distribution of facies and environments in an alluvial fan (after McGowan and Groat, 1971) (subfacies of the GT facies)

competence decreased in the distal reaches of the fan. The Sp subfacies was formed by the migration of transverse bars in shallow water under low flow regime conditions (Smith, 1970; Hein and Walker, 1977). The Sh and Sl subfacies, in contrast, were deposited from shallow upper regime stream flow as horizontal strata and shallow scour fills (Rust, 1978). The fine grained subfacies of the GS facies, Fl and Fm, also formed on the distal part of the fan primarily as overbank deposits. The vertebrate fossils which occur in thin conglomerates (Gm subfacies), conglomeritic sandstones (Gt subfacies), and mudrocks (Fl subfacies) of the GS facies were preserved in the distal part of the fan system. Tektites were also deposited on the distal fan in Gm and Gt subfacies.

Miall (1978) recently suggested that most braided stream deposits may be classified by vertical profile models. The various subfacies of the GS facies are representative of the Scott, Donjek, and South Saskatchewan profile types (Miall, 1978) which form a gradational proximal-distal sequence in the central Cagayan Valley. This sequence reflects a downstream decrease in the gravel/sand ratio from the Scott type (>90% gravel) to the Donjek (10-90% gravel) and South Saskatchewan (<10% gravel) types.

The interpretation that the braided stream deposits of the GS facies represent a Pleistocene alluvial fan is based on the distribution of subfacies and the interfingering relationship with the GT facies. The subfacies of the GS facies reflect proximal-distal variations of braided streams within a relatively small area (45 km square). McGowan and Groat (1971) have described similar proximal-distal facies changes in a fan formed by the Van Horn Sandstone, Texas. In this fan, the proximal framework conglomerates (GM facies) grade to distal finer grained cross bedded sandstones and fine grained sediments in a distance of 30 to 40 km. Braided streams on alluvial plains, in contrast, are characterized by more gradual proximal-distal facies changes (Rust, 1978). In the Donjek River, for example, gravel is still the dominant type of sediment 50 km from the river's source (Rust, 1972). The rapid change from proximal to distal facies in the central Cagayan Valley, therefore, suggests that the GS facies was deposited as an alluvial fan. The greater thickness and coarser grain size of the GS facies in the south at an equal distance from the mountain front in contrast to the north reflects the cone shaped morphology of the ancient fan system. The coarse grained conglomerates in the south indicate that the apex of the fan was near the southwest part of the study area. The finer grained subfacies of the northern and eastern part of the study area were formed in the arcuate distal portion of the fan which extended around the fan apex. The GT facies which interfingers with the braided stream deposits of the GS facies contains a Gms subfacies which is interpreted as debris flow deposits (next section). Debris flow deposits are rarely preserved in alluvial plain braided stream environments but are characteristic deposits of alluvial fans (Rust, 1978).

The alluvial fan formed primarily as a result of tectonic activity. Increased uplift of the Cordillera Central during the Pleistocene resulted in progradation of the fan into the Cagayan basin. The progradation of the fan is reflected by the coarsening upward of the GS facies in the Awidon Mesa Formation. Multiple coarsening upward or fining upward sequences which are characteristic of some fans (Collinson, 1978a) were not observed in the coarser proximal detritus. The abrupt and unordered vertical textural changes in the facies, instead, suggest that there was frequent shifting of shallow channels and bars (Harms et al., 1975). Some fining upward sequences occur in the distal deposits and probably reflect the transition to higher sinuosity streams.

Massive Matrix-Supported Pebble to Boulder Conglomerate, Tuff-Breccia, and Tuff (GT) Facies

Description

The GT facies occurs throughout the Awidon Mesa Formation increasing in thickness in the upper part and along the western portion of the Central Cagayan Valley. The facies is commonly 1 to 3 meters thick along the flanks of Cabalwan, Enrile, and Pangul anticlines where it interfingers with the GS and GSF facies. It is generally thicker, up to 20 m, along the west flank of Enrile Anticline. The facies is well-exposed along the Tabuk plateau where it is up to 25 m thick. From the Tabuk Plateau, it extends discontinuously into the Cordillera Central where a thickness of 300 m has been recorded by Durkee and Pederson (1961) at Awidon Mesa.

The facies is composed of light gray (N7) to white (N9) pyroclastic deposits which are divided into 5 subfacies as summarized in Figure 50. The coarser grained subfacies, Tms, Gms, and Tt, occur only in the Awidon Mesa Formation whereas the finer grained well sorted subfacies, Ts and Tr, occur in both the Awidon Mesa Formation and the Upper Member of the Ilagan Formation.

The Tms, Gms, and Tt subfacies commonly occur in distinct vertical sequences which form pyroclastic complexes up to 25 m thick in the Awidon Mesa Formation. The Tms subfacies, massive matrix supported tuff breccia, is the dominant subfacies of the pyroclastic complexes. It occurs as laterally continuous massive beds (Figure 51a and b) up to 8 m thick which display reverse grading of all pyroclasts or reverse grading of pumice and normal grading of lithic pyroclasts. The reverse grading of all pyroclasts is typical of the Tms subfacies in the Tabuk Pyroclastics (Figure 51c)

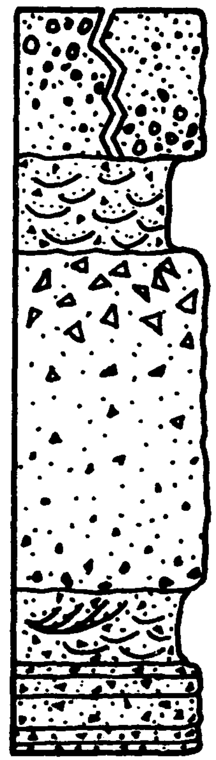
SUBFACIES	FACIES CODE	VERTICAL SEQUENCE	SEDIMENTARY STRUCTURES	THICKNESS	INTERPRETATION
Gravel, massive matrix-supported	Gms		None, normal or reverse grading	5 m	Lahars (including both mud flows and debris flows)
Tuff, well sorted	Tr		Small scale trough cross beds	cm-2 m	Reworked tuff
Tuff-breccia, massive, matrix-supported	Tms		Normal grading of lithics, reverse grading of pumice, reverse grading of all pyroclasts in some deposits gas escape structures	8 m (single flow unit)	Pyroclast flow (ignimbrite)
Tuff, fine to coarse grained	Tt		Trough cross beds	cm-2 m	Ground surge deposit
Tuff, well sorted	Ts		Stratified	cm-2 m	Pyroclast fall (airfall tuff)

Figure 50. Subfacies of the massive matrix-supported pebble to boulder conglomerate, tuff-breccia and tuff (GT) facies (the occurrence of the tuffs varies in the sequence)

Figure 51. Photographs of tuff-breccias (ignimbrites) of the massive matrix-supported pebble to boulder conglomerate, tuff-breccia, and tuff (GT) facies

a. Laterally continuous tuff-breccia of the Tms subfacies (ignimbrite) deposited as one flow unit, northern Pangul Anticline

b. Tms subfacies (ignimbrite) composed of numerous flow units, Liwan Pyroclastic Complex, east flank of Enrile Anticline

c. Tuff-breccia of the Tms subfacies (ignimbrite) with crystal rich base and subangular dacite clasts which coarsen upward, Tabuk Pyroclastics, Tabuk plateau

d. Tms subfacies (ignimbrite) with gas escape structures which overlies a tuff and claystone, Minabbag

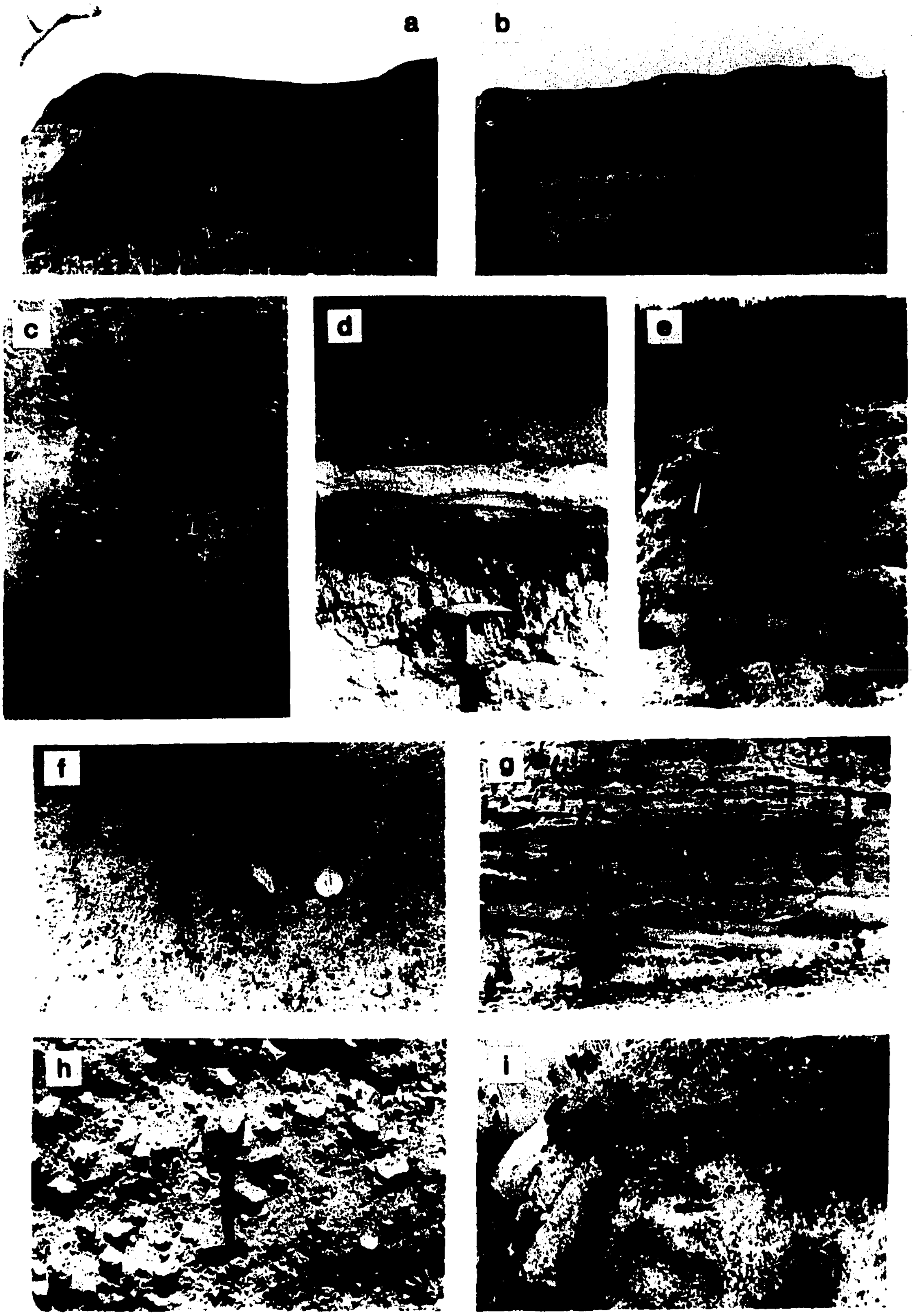
e. Elongate gas escape structures as seen along upper surface of the Tms subfacies (ignimbrite), southern Pangul Anticline

f. Wavy gas escape structures in the Tms subfacies (ignimbrite), Liwan Pyroclastic Complex, west flank of Enrile Anticline

g. Trough cross bedded Tt subfacies (ground surge deposit), Tabuk Pyroclastics, Tabuk plateau

h. Tuff-breccia of the Tms subfacies (ignimbrite) with subangular dacite clasts, Tabuk Pyroclastics, Tabuk plateau

i. Tuff-breccia of the Tms subfacies (ignimbrite) with reverse grading of rounded pumice clasts, northern Pangul Anticline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whereas normal grading of lithics and reverse grading of pumice (Figure 51i) is characteristic of the subfacies in the Liwan Pyroclastic Complex. The lithic pyroclasts, which are primarily subangular andesite and dacite (Figure 51h) up to boulder size, commonly float in a dacitic tuff matrix. The pumice fragments, in addition to coarsening upward, are often concentrated along inversely graded layers. The basal contact of the subfacies is always sharp and usually planar (Figure 51d) while the basal portion of the subfacies is characteristically crystal rich and finer grained than the rest of the subfacies (Figure 50c). Gas escape structures, vertical pipe-like structures which lack the finer grained material of the deposit, commonly occur in the upper portion of the facies (Figures 51d, e, and f). These structures are typically 1 cm wide and wavy as seen in side view (Figure 51f). As observed along bedding planes (Figure 51e), some of the gas escape structures are elongate. The geometry of the subfacies varies with the age of the deposit. The younger deposits occur as large sheets which are traceable for up to 15 km. The older deposits, in contrast, are lenticular or grade laterally into the GS facies. The Tt subfacies is formed by small to large scale trough cross bedded fine to coarse grained dacitic tuff (Figure 51g) which occurs stratigraphically beneath numerous outcrops of the Tms subfacies. The subfacies varies from several centimeters to 2 m in thickness. In most outcrops, the Tt subfacies is relatively thin, 10-20 cm, and the cross bedding is very faint. The Gms subfacies, massive matrix-supported conglomerates (Figure 52a) commonly overlies the Tms subfacies. This subfacies also displays reverse grading (Figure 52b) or normal grading (Figure 52c), but the composition and texture are usually more varied than the Tms subfacies. Subrounded

Figure 52. Photographs of conglomerates and tuff-breccies (lahars) and tuffs of the massive matrix-supported pebble to boulder conglomerate, tuff-breccia, and tuff (GT) facies

a. Conglomerate of the Gms subfacies (lahar) which contains subrounded pebble to boulder size andesite, dacite, and basalt, Tabuk Pyroclastics, Tabuk plateau

b. Tuff-breccia of the Gms subfacies (lahar) composed of subangular dacite and andesite, Tabuk Pyroclastics, Tabuk plateau

c. Conglomerate of the Gms subfacies (lahar) with normal grading of pebble to boulder size dacite and andesite clasts, Tabuk Pyroclastics, Tabuk plateau

d. Thinly laminated to thin bedded tuff, ts subfacies (air fall tuff), Liwan Pyroclastic Complex, west flank of Cabalwan Anticline

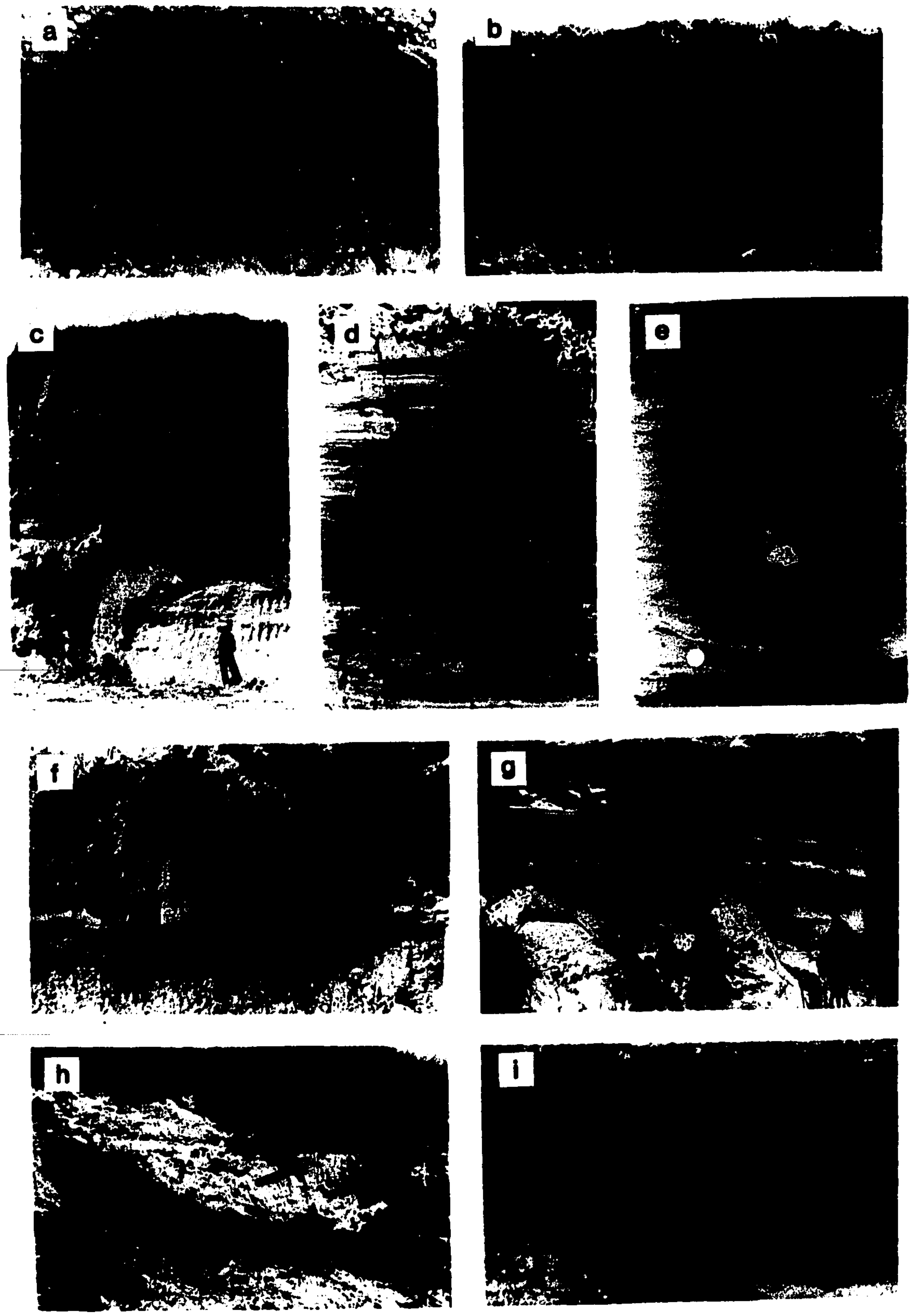
e. Tuff of the Tr subfacies (reworked) with small scale trough cross bedding, Tabuk Pyroclastics, Tabuk plateau

f. Tuff of the Ts subfacies (air fall tuff) overlying flood plain claystone and overlain by an ignimbrite (Tms subfacies) with a fine grained crystal rich base, Liwan Pyroclastic Complex, west flank Enrile Anticline

g. Thinly bedded vitric tuff of the Ts subfacies (air fall tuff), Liwan Pyroclastic Complex, west flank, Enrile Anticline

h. Lenticular massive tuff of the ts subfacies, east flank, Cabalwan Anticline

i. Thin tuff beds deformed by loading, Espinosa Ranch, Pangul Anticline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andesite and dacite clasts predominate in some deposts (Figure 52a) in addition to subangular clasts. Subrounded to subangular basalt clasts also occur but are usually only pebble to cobble size in contrast to the cobble to boulder size clasts of andesite and dacite. The orientation of the clasts varies in the subfacies from clasts oriented parallel to the bedding or imbricated to randomly oriented clasts. The matrix which commonly forms more than half the subfacies is composed of poorly sorted dacitic tuff. Most deposits of the facies have sharp planar basal contacts, but irregular scoured contacts also occur. The subfacies, which is up to 5 m thick, is commonly distinguished from the Tms subfacies by the lack of a fine grained crystal-rich base and gas escape structures in addition to compositional and textural differences.

The Ts and Tr subfacies are both composed of well-sorted tuff (Figure 50). The Ts subfacies is characterized by well-stratified laminae (Figure 52d) to thin bedding (Figure 52g) or may appear massive. The Tr subfacies, in contrast, displays small scale trough cross bedding (Figure 52e) and concentrations of heavy minerals along bedding surfaces. The basal contacts of both subfacies are usually sharp and planar, but the upper contacts are commonly deformed by loading of the overlying bed (Figure 52f and i). Both subfacies are lenticular (Figure 52h) with a maximum thickness of 2 m in the Awidon Mesa Formation. These tuffs, which are interbedded with the GS facies and other subfacies of the GT facies, commonly pinch out over short distances. More laterally extensive tabular tuffs, primarily the Sr subfacies, occur in the Awidon Mesa Formation where the subfacies forms tuffaceous intervals up to 10 m thick which interfinger with the GSF facies.

Depositional environment

The subfacies of the GT facies are interpreted as nonwelded or sillar ignimbrites (Figure 51), lahars, and tuffs (Figure 52). Some deposits represent individual ignimbrites or lahars as in Figures 51a and 52c, but several subfacies often occur together forming pyroclastic complexes. In these complexes, several ignimbrites may occur in a sequence, or individual ignimbrites may be interbedded with tuffs and lahars or grade upward to lahars. The tuffs commonly occur as individual units interbedded in the GSF and GS facies as well as the GT facies.

The term ignimbrite is used as defined by Sparks et al. (1973) and Sparks (1976) to refer to pyroclast flow deposits of Peléan eruptions which are characterized by three distinct layers in a vertical sequence. These are (1) a lower ground surge deposit, (2) the main ignimbrite flow unit, and (3) a fine ash deposit (Figure 53). The cross bedded deposits of

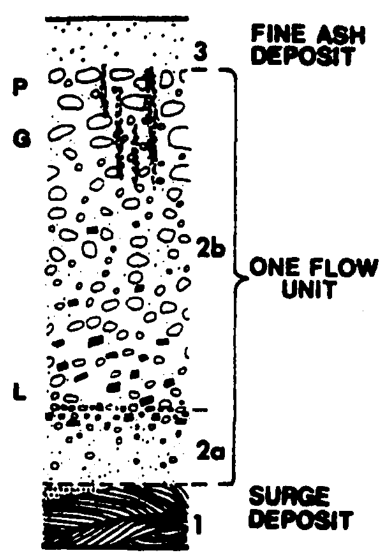

Figure 53. Schematic section of pyroclastic deposits from a Peléan eruption (L = lithics, P = pumice, G = gas escape structures (after Sparks et al., 1973))

the Tt subfacies are interpreted to be ground surge deposits (Sparks and Walker, 1973) which preceded the main flow. In deposits with many flow units, some of the surge layers may have been deposited at the top of pyroclast flow sequences as recently described by Fisher (1979). The main flow units of the Cagayan Valley ignimbrites are massive matrix supported tuff-breccias of the Tms subfacies. They have a diagnostic crystal rich basal layer (Walker, 1971; Sparks et al., 1973) which is finer than the rest of the unit which is characterized by a normal or reverse grading of lithic clasts and a reverse grading of large pumice clasts. Gas escape structures (fossil fumeroles of Walker, 1971) are preserved in many deposits and were formed as gas escaped from the settling flow. The elongate gas escape structures observed in one deposit probably formed while the flow was still moving. The fine ash deposit or coignimbrite ash (Sparks and Walker, 1977) is a thin, very fine ash deposit of the Ts subfacies which, in a complete sequence, overlies the main ignimbrite flow unit. It is not often preserved in ignimbrites composed of only one flow unit.

The ignimbrites indicate that both Plinian and Peléan eruptions occurred during the Pleistocene. These eruptions probably took place in the Cordillera Central, the closest possible source. Valley filling pyroclastic deposits occur in the Cordillera at Awidon Mesa and several mesa-like localities further to the west. Short duration, highly explosive Plinian eruptions preceded the Peléan eruptions. This is indicated by the abundance of pumice clasts in the ignimbrites which typically originate during an initial Plinian phase of volcanism which produces pumice fall deposits (Sparks et al., 1973). The ignimbrites were generated by subsequent Peléan eruptions of dacitic magma which were of longer duration. Williams and McBirney (1979) note that the eruptive cycle of Peléan eruptions generally lasts a few years. The layers in the ignimbrite deposits were formed by differentiation of the material that was erupted (Fisher, 1979). The ground surge was formed by collapse of the outer part of the eruption column. Progressive collapse of the eruption column interior produced the voluminous high concentration pyroclast flows which, according to Sparks (1976), are commonly laminar in their movement. The coignimbrite ash was deposited from an ash cloud that was segregated from the surface of the flow. According to Fisher (1979), it may be overlain by a thin fallout deposit with characteristics of a surge deposit.

The variations in geometry of the ignimbrites are interpreted to reflect proximal-distal variations. The older discontinuous lenticular deposits which help define the base of the Awidon Mesa Formation are interpreted as distal ignimbrites. These grade laterally to cross bedded dacitic conglomerates and sandstones within several hundred meters. The younger extensive sheet-like deposits are interpreted as more proximal ignimbrites.

Criteria for the distinction between distal and proximal ignimbrites (Table 15) have recently been described by Sheridan (1979). According to these criteria, the ignimbrites of the Awidon Mesa Formation are primarily distal ignimbrites. The ignimbrites of the Liwan Pyroclastic Complex are characterized by fine grained laminar shear layers at the base, strong concentrations of pumice at the top, few large lithic clasts, inversly graded layers capped by pumice concentrations, and fine grained air fall tuffs which cap the ignimbrite. Some characteristics of proximal

Table 15. Characteristics of proximal and distal ignimbrites (from Sheridan, 1979)

Proximal ignimbrites	Distal ignimbrites
1. Plinian pumice fall deposit at base	1. Finer size plinian pumice fall deposits at base
2. Thick ground surge deposit	2. Surge deposits are lacking
3. Basal shear zone lacking	3. Laminar shear layer is usually present at base
4. Pumice fragments are nongraded or normally graded	4. Strong concentrations of pumice at top
5. Gas escape structures are common	5. Flow units have few large lithic pyroclasts
	6. Contains numerous inversly graded layers with pumice concentrations at top
	7. Fine grained air fall tuff caps sequence

ignimbrites, surge deposits, and gas escape structures also occur, but the overall characteristics are more representative of distal ignimbrites. The ignimbrites of the Tabuk plateau contain primarily large lithic pyroclasts and the thickest surge deposits, indicating that the Tabuk Pyroclastics are more proximal to the source than the Liwan Pyroclastic Complex.

The massive matrix-supported conglomerates of the Gms subfacies are interpreted as lahars. Lahar is an Indonesian word for volcanic breccias which were transported down the slopes of a volcano by water (van Bemmelen, 1949). As suggested by Crandell (1971) and Williams and McBirney (1979), lahar is here used as a general term for both volcanic mudflows and debris flows. It is pertinent to note, however, that the Gms subfacies contains both debris flows and mudflows. Mudflows, which contain at least 50% sand, silt, and clay (Varnes, 1958), are the most common type of lahar while debris flows, which contain less than 50% sand, silt, and clay, are restricted to the western side of the study area.

As Crandall (1971) notes, there is no single feature which may be used to distinguish lahars from all other kinds of coarse deposits. Some lahars are normally graded (Mullineaux and Crandell, 1962) while others have been described as reversely graded (Schmincke, 1967). Considering flow dynamics, Hampton (1975) suggested that debris flows may be normally or reversely graded. Clast orientation varies between lahars as clasts may be oriented parallel to the bedding (Fisher, 1971; Enos, 1976) or lack preferred orientation (Reineck and Singh, 1975). The character of the basal contact also varies as the base of the underlying deposit may be scoured (Schminke, 1967) or lack scour features (Enos, 1976). Despite these variations, which also occur in the Cagayan Valley lahars, there are several distinguishing features which make it possible to identify the lahars. These are the massive, poorly sorted, matrix-supported character of the Gms subfacies and the occurrence of subrounded to subangular clasts of various lithologies. The variations in clast composition and rounding enables lahars to be differentiated from ignimbrites as ignimbrites contain only subangular andesite-dacite or rounded pumice pyroclasts. Lahars also lack the associated ground surge deposit, the upper fine ash deposit, and the gas escape structures which are diagnostic of ignimbrites.

The lahars originated as pyroclast flows which were mobilized by water from rains during or after volcanic eruptions. The water soaked pyroclastic debris flowed from the Cordillera Central along drainage ways and then spread out into the Cagayan Valley along stream channels and floodplains. Accretionary lapilli, which commonly form in ash clouds when it rains, have not as of yet been found in the Cagayan Valley pyroclastic deposits.

The well-sorted tuffs of the Ts and Tr subfacies are pyroclast fall deposits. The good sorting and stratification of the Ts subfacies indicates that it represents air fall tuffs. Mantle bedding, a diagnostic characteristic of air fall tuffs (Sparks and Walker, 1973), was not observed, however, due to limited outcrops. The tuffs of the Tr subfacies are interpreted to be reworked by water or wind as indicated by the trough cross bedding and heavy mineral concentrations.

The GT facies coarsens upward from tuffs in the Upper Member of the Ilagan Formation to ignimbrites and lahars in the Awidon Mesa Formation. This coarsening upward of the facies reflects the migration of volcanic centers in the Cordillera to the east (away from the trench system) with increasing maturity of the arc system and uplift of the Cordillera. In the Plio-Pleistocene, only the volcanic ash was transported to the central Cagayan basin. In the Pleistocene, coarser pyroclast flows were transported to the basin from a more proximal source in the Cordillera. The interfingering of the Tms and Gms subfacies with the GS facies indicates that the ignimbrites spread out over an alluvial fan which also prograded eastward through the Pleistocene in response to uplift of the Cordillera.

PROVENANCE

The provenance of the central Cagayan Valley Plio-Pleistocene sediments is summarized diagrammatically in Figure 54. The composition of the sediments indicates that they were derived from volcanic and plutonic igneous source rocks in the adjacent volcanic arcs. Most sediments, especially the pyroclastic deposits, are probably first cycle, but some were undoubtably reworked from the eroded Miocene sediments along the foothills of the arcs. The close proximity of the source rocks is indicated by the textural immaturity of the poorly sorted sandstones and the abundance of unstable heavy minerals.

Compositional variations and lithofacies distribution indicate that the importance of various arcs as source areas changed in the Plio-Pleistocene. The Ilagan Formation sandstone, feldspathic litharenites, and litharenites which are characterized by a pyroxene heavy mineral association were derived from the arcs to the east and south. The Cordillera Central to the west contains similar source rocks and also probably contributed detritus to the Ilagan Formation. Due to the similarity of the source rocks, individual sources of the Plio-Pleistocene sediments cannot be differentiated. The Ilagan Formation contains three facies which record the progradation of a delta system and the flow of a large meandering stream to the north. The distribution of these facies suggests that the Ilagan Formation sediments were derived from the volcanic arcs bordering the southern portion of the basin. The Cordillera Central became the principal source of sediments in the Pleistocene. The percentages of quartz and feldspar are greater in the Awidon Mesa Formation sandstones,

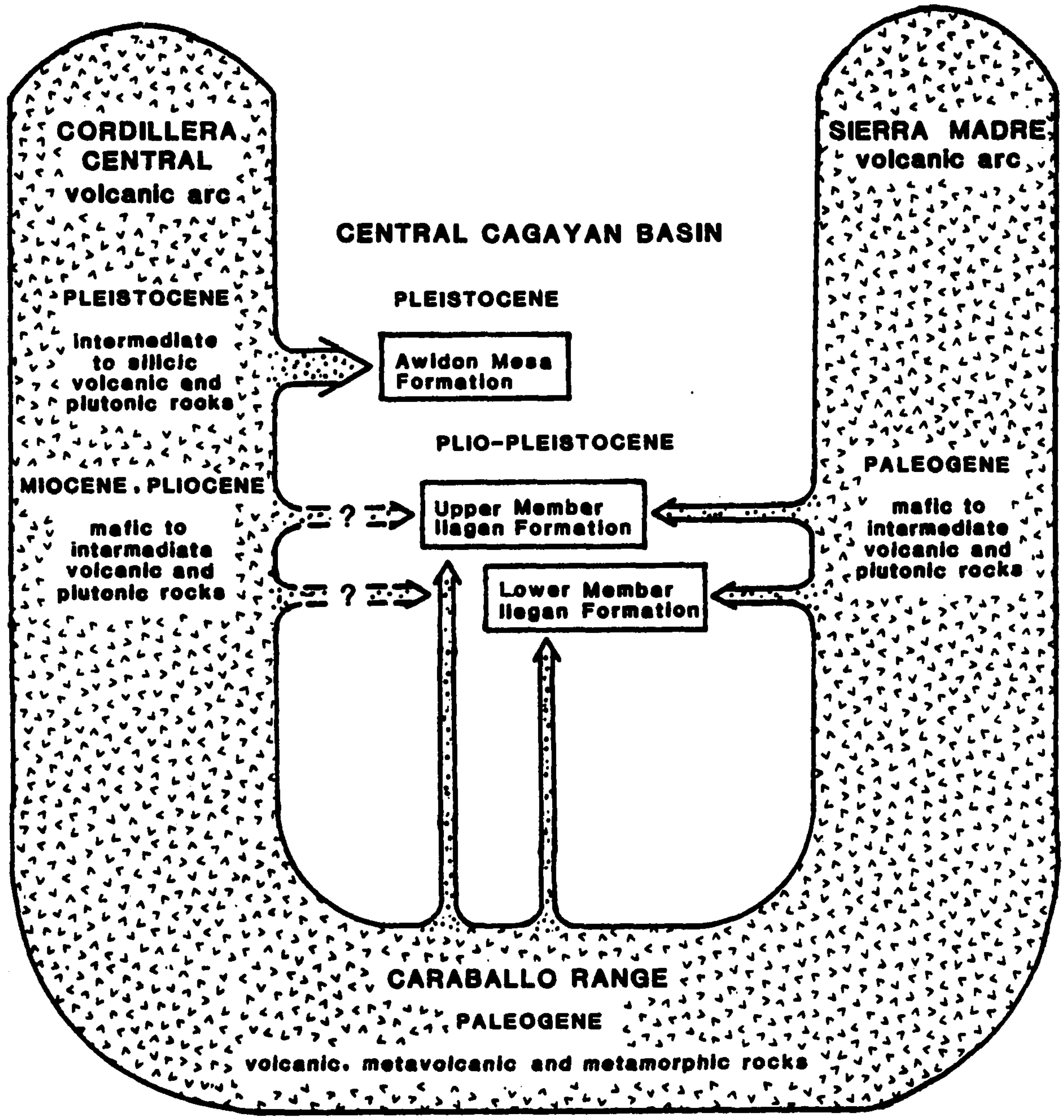

Figure 54. Provenance of the Plio-Pleistocene sediments, central Cagayan basin

lithic arkoses, which are characterized by hornblende and oxyhornblende heavy mineral associations. The increase in quartz and feldspar and the abundance of hornblende and oxyhornblende reflect the intermediate to silicic volcanic activity in the Cordillera Central during the Pleistocene. The ignimbrites and lahars of the Awidon Mesa Formation also reflect the contribution of volcanic sources in the Cordillera. These sources are probably located west of Awidon Mesa. Flat mesa-like areas west of Awidon Mesa are distinctive on topographic maps of the Cordillera Central and are probably upheld by ignimbrites similar to those at Awidon Mesa. Durkee and Pederson (1961) suspect that four little known volcanic vents that occur in this area (Alvir, 1956) supplied or are related to vents which supplied the pyroclastic material to the Awidon Mesa Formation. Aside from the compositional data, the coarsening upward of the Pleistocene lithofacies reflects the increasing contribution from the Cordillera Central in the Pleistocene.

PALEOGEOGRAPHY AND PALEOENVIRONMENTS

The Cagayan basin began to form in the late Oligocene to early Miocene epochs following the polarity reversal of the Luzon arc system (Figure 5) and the initial uplift of the ancestral Cordillera Central (Durkee and Pederson, 1961; Karig, 1973; DeBoer et al., 1980). Following Miocene marine sedimentation, regional uplift with very little tilting or compressional deformation occurred (Christian, 1964) resulting in the transition from marine to terrestrial sedimentation.

Paleogeographic reconstructions and paleoenvironmental interpretations of the central Cagayan Valley Plio-Pleistocene deposits are based primarily on the distribution of major lithofacies. The vertical sequence and lateral distribution of facies documents a regression of the Pliocene sea and the filling of the basin with detritus from the surrounding volcanic arcs. Compositional variations, fossil assemblages, and the timing of tectonic events provide additional data for paleogeographic and paleoenvironmental interpretations. Most dates used in the following discussion are based on previous investigations because radiometric dating of samples collected during this study has not yet been completed.

The beginning of the Pliocene epoch is marked by the transition from marine to deltaic sedimentation as illustrated in Figure 55. Lower Ilagan delta front sediments of the interbedded fine grained sandstone and mudstone (SF) facies were deposited on Miocene prodelta clays of the underlying Baliwag and Buluan Formations. Thinly interbedded fine grained sandstones, siltstones, and mudstones formed a distal bar on the seaward margin of the advancing delta front. Some of the deformed bedding observed in the thinly interbedded sandstones and siltstones may represent

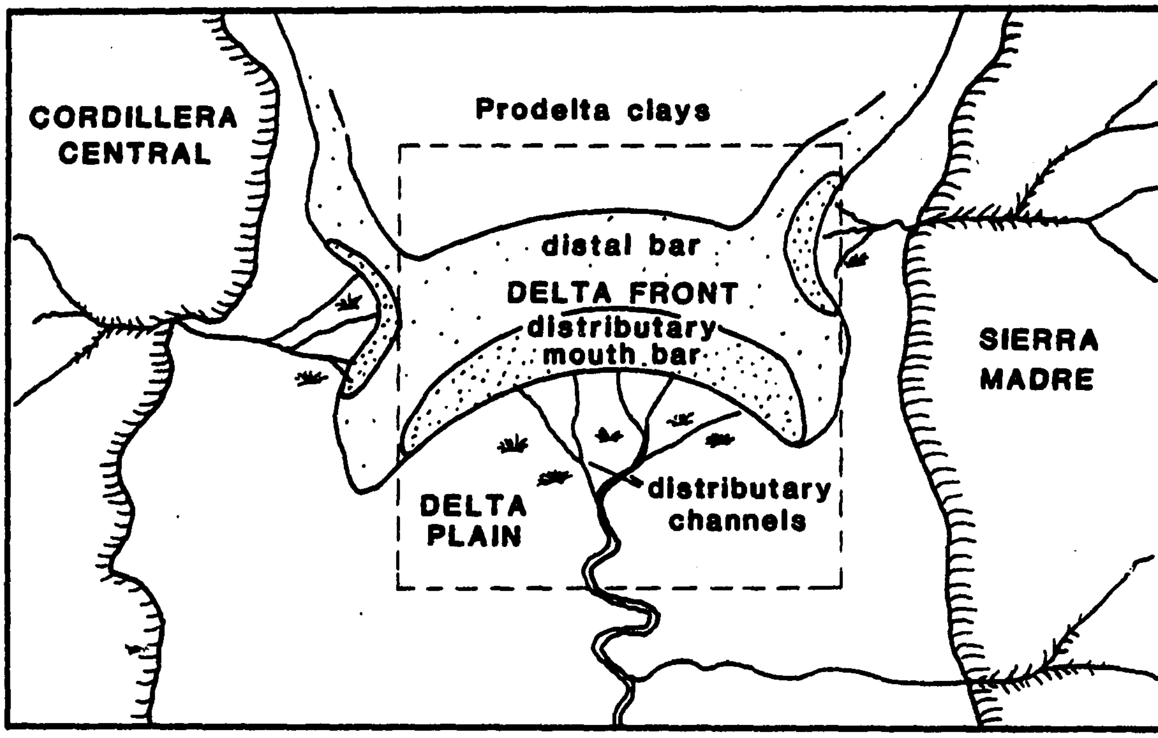

Figure 55. Schematic representation of the central Cagayan basin paleogeography during deposition of the Lower Member of the Ilagan Formation (Pliocene)

penecontemporaneous slumping as the delta subsided and advanced. Shell remains are common in the sandstones and were transported by higher energy currents which deposited the thin sand beds. Small burrowing organisms lived along the distal bar as indicated by the occurrence of burrows up to 2.5 cm in diameter. Sediment was transported to the delta front distributary mouth bar by subaqueous and subareal distributary channels of the lenticular, cross bedded, medium grained sandstone and siltstone facies. The distributary channels flowed across a delta plain composed of finer grained sediments of the facies which were deposited during storms and periods of flood. These finer grained sediments were bioturbated to varying degrees by plants and burrowing animals. The delta front and delta plain sediments of the Lower Member of the Ilagan Formation were deposited on at least one lobate river dominated delta system which prograded to the north. Wave and tidal influences did not have as great an effect on delta growth as did the constant influx of sediment by a river. A major delta must have been formed by the deposition of sediments transported from the southern part of the basin. Minor deltas were also probably formed by rivers draining the Sierra Madre and Cordillera Central.

At least one major meandering stream flowed to the north (Figure 56) across the central Cagayan Valley in the Plio-Pleistocene and deposited 500 m of the Upper Ilagan polymictic conglomerate, trough cross bedded sandstone, and claystone (GMF) facies. The river flowed in a channel up to

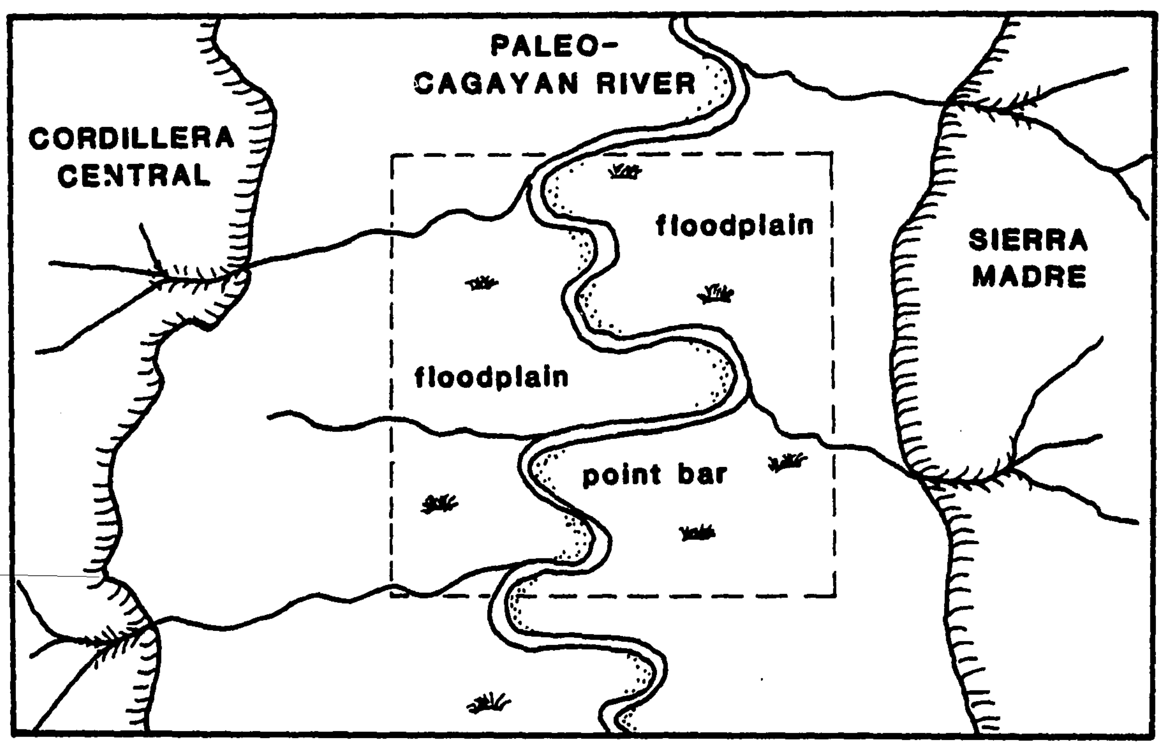

Figure 56. Schematic representation of the central Cagayan Basin paleogeography during deposition of the Upper Member of the Ilagan Formation (Plio-Pleistocene)

12 m deep as indicated by the thickness of the conglomerates and sandstones. Changes in conglomerate thickness and clast size indicate the increasing contribution of a westerly source as the basin filled and the Cordillera Central was uplifted. The older conglomerates of the Upper Member of the Ilagan Formation are thin and composed of granules and pebbles which were transported primarily from the southern part of the Cagayan Valley by a major axial stream or from the Cordillera Central which was then a distant westerly source. The conglomerates thicken and coarsen to pebble-cobble conglomerates in the upper part of the Upper Member reflecting the increasing contribution of a more proximal westerly source, the Cordillera Central volcanic arc, which was active during the Pliocene. Volcanic ash was reworked by streams and deposited in association with sandstones, siltstones, and claystones in the swales of point bars and on the floodplains where soils developed as indicated by the numerous horizons with blocky structure and several iron rich pisolitic paleosols. Permineralized logs up to several meters long indicate that trees were eroded, transported, and deposited by the stream. Root casts and leaf impressions in tuffs indicate a variety of plant cover existed in the area.

During the Pleistocene, tectonic and volcanic events in the Cordillera Central greatly influenced the sedimentary history and paleogeography of the central Cagayan basin. Geanticlinal uplift of the Cordillera Central and eastward migration of volcanic centers resulted in a migration of the basin axis to the east and the formation of an alluvial fan complex along the faulted mountain front (Figure 57). Coarse clastics of the clast-supported polymictic conglomerate and sandstone (GS) facies were transported to the valley by braided streams. The coarser sediments were

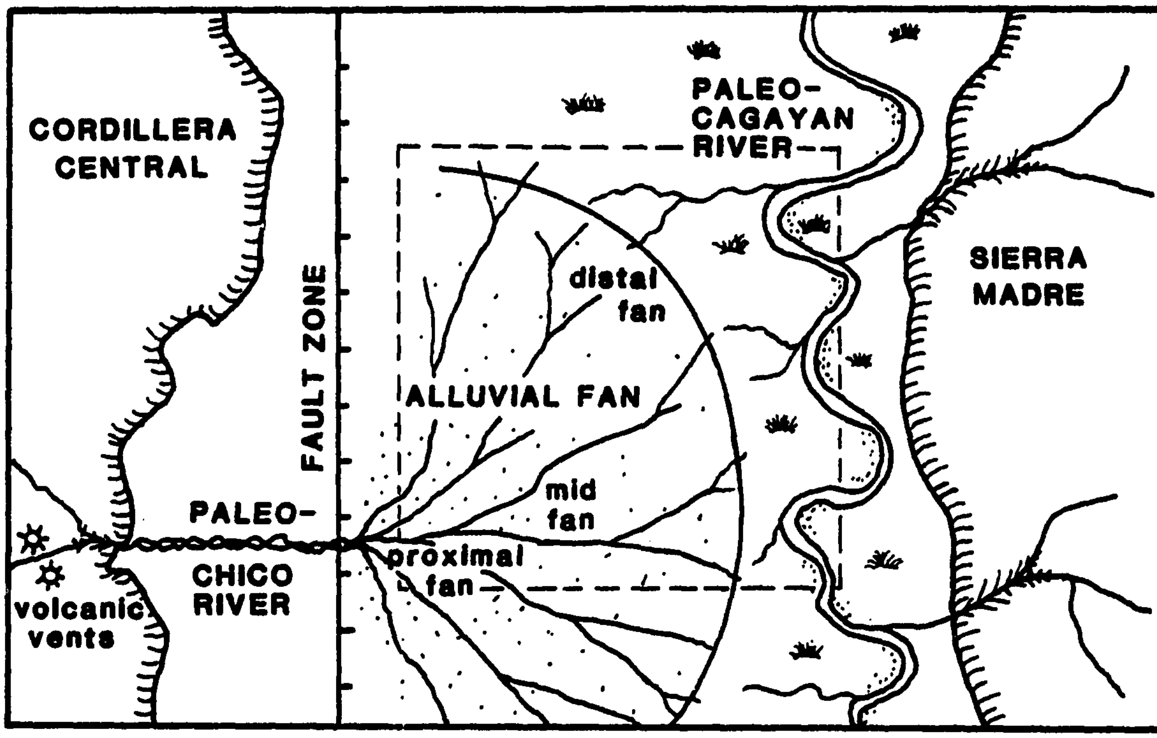

Figure 57. Schematic representation of the central Cagayan basin paleogeography during deposition of the Awidon Mesa Formation (Pleistocene, before folding)

deposited on the inner fan in the southwestern part of the study area. The finer sediment was deposited by distal braided streams on the toe of the fan or transported by small meandering streams to the axial meandering stream which probably flowed along the eastern side of the valley where paleo-Cagayan River terraces now occur. As uplift of the Cordillera continued, proximal braided stream deposits developed farther out in the basin covering the distal deposits (Figure 58a). The alluvial fan probably had a relatively low slope. The slopes of alluvial fans are relatively gentle in humid areas because the flow of water shifts the sediment and decreases the gradient (Bull, 1964).

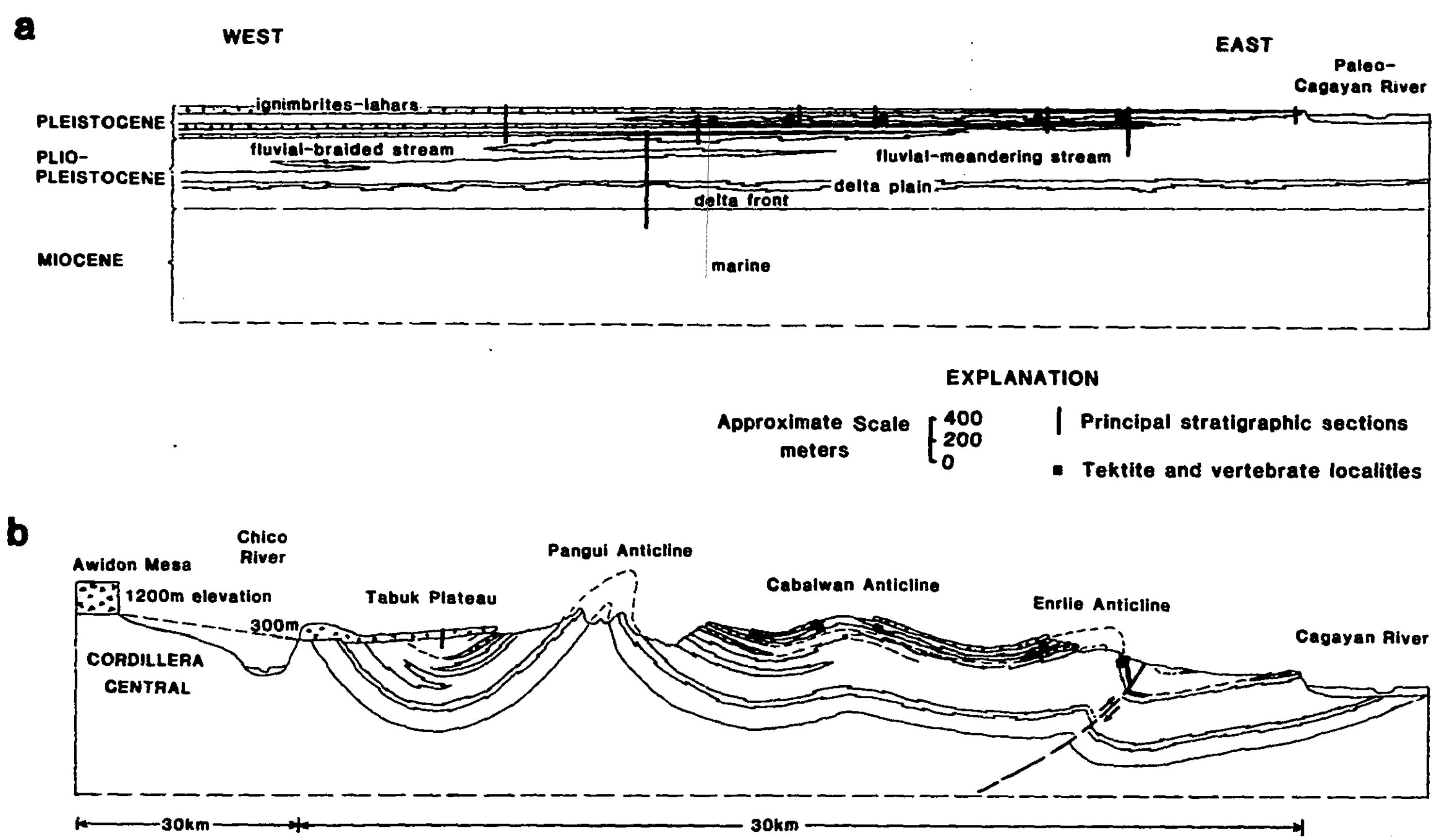

Figure 58. Interpretive cross sections of the central Cagayan Valley before (a) and after (b) middle Pleistocene folding

Voluminous pyroclast flows were produced in the Cordillera Central during the Pleistocene by explosive plinian and peléan volcanic activity. The pyroclast flows and detritus mobilized by water, lahars, flowed along topographic depressions from the Cordillera Central to the valley where they spread out on the alluvial fan and alluvial plain of the paleo-Cagayan River. The dacitic composition and quartz phenocrysts in the pyroclast flow deposits indicate that volcanism became more silicic in the Pleistocene in contrast to the mafic to intermediate volcanism of the Miocene and Pliocene and suggests that the Cordillera volcanic arc has matured. Ragland et al. (1976) and DeBoer et al. (1980) describe three volcanic belts in western Luzon, a tholeiitic belt in the west, a calc-alkaline belt, and a high-K calc-alkaline belt in the east. The high-K calc-alkaline belt reflects the maturity of the arc as high-K calc-alkaline volcanism occurs only during the advanced stages of island arc evolution (Gill and Gorton, 1973).

A terrestrial vertebrate fauna migrated to the Cagayan Valley via at least one land-bridge during the middle Pleistocene. A land-bridge between Luzon and Taiwan is suggested by faunal and geographic considerations but cannot be proved at this time. When the vertebrates arrived, the valley contained a large north flowing meandering stream with an extensive flood plain and an alluvial fan complex forming along the Cordillera Central mountain front. All the vertebrate fossils found *in situ* were transported and buried in the coarse deposits or floodplains of small meandering streams or distal braided streams between the toe of the alluvial fan and the paleo-Cagayan River (Figures 57 and 58). This area probably supported a favorable habitat for the fauna as well as favorable conditions for fossilization. Reineck and Singh (1975) note that sediments of alluvial fans are deposited in extremely oxidizing conditions and that organic and fossil remains are rather rare. Preservation of the Cagayan Valley fauna probably occurred because of the greater percentage of fine grained sediment in the distal alluvial fan and meandering stream deposits. The absence of vertebrate fossils in the southwestern part of the study area is probably the result of the poor preservation potential of the more proximal alluvial fan deposits.

Tektites of possible extraterrestrial or impact origin were deposited by distal braided streams and/or small meandering streams on the plain between the alluvial fan and the paleo-Cagayan River (Figure 12). The tektite date (0.92 $\pm$ 0.17 Myr B.P.), limited stratigraphic occurrence of tektites and the association with middle Pleistocene fauna suggest that they are middle Pleistocene in age and probably correlative with the 710,000 year B.P. tektites of Java (von Koenigswald, 1967). Among other theories, Dietz and McHone (1976) have recently suggested that the 710,000 year B.P. tektites of the Southeast Asian strewnfield may be products of the meteoritic impact at El'gytgyn, Siberia.

Little can be inferred regarding Plio-Pleistocene climates from compositional and textural characteristics of the sediments. It has been suggested that feldspar alteration (Todd, 1978) and size-composition studies (Young et al., 1975; Mack and Suttner, 1977) can be used to interpret paleoclimates. The extensive diagenetic alteration of the Plio-Pleistocene Cagayan basin sediments has made it impossible to make paleoclimate interpretations based on framework grain composition and texture. Walker (1978) and James et al. (1981) have noted that diagenetic alteration of the sediments commonly limits the paleoclimate interpretations that can be made.

The Pleistocene climate of the Cagayan Valley can be inferred from a regional synthesis of Pleistocene climatic variations in Southeast Asia compiled by Verstappen (1975). He relates changes in sea level, temperature, and, more importantly, rainfall and humidity changes to changes in vegetative cover, soil formation, and landform development (Table 16). This information is important to consider in cultural interpretations of Pleistocene archaeologic sites (assuming the Cagayan Valley paleolithic tools can be found _in situ_) because early human culture reflects adaptations to environments. Based on Verstappen's (1975) summary and field work, a tree savannah environment probably predominated in the Cagayan Valley

Table 16. Summary of Pleistocene climatic interpretations for Southeast Asia (from Verstappen, 1975)

1. Sea level dropped 100 m converting 3,000,000 km^3 of shallow water seas to land; sea water temperatures dropped 4-5^{o}C.
2. 30% drop in rainfall below present values during glacial periods; distribution of rainfall different.
3. Drop in rainfall produced a more pronounced dry season which caused drought stress in vegetation and changes in soils, fauna, and geomorphological processes.
4. Botanical evidence suggests drier conditions have been the exception in the past; humid tropical conditions are thought to be more normal. Aridity was never so severe that steppe or grass savannah could become established. Monsoon forest and tree savannah were probably the characteristic vertebrate environments during the glacials.
5. Humidity increases may have been responsible for the extinction of the grazing mammals.

during the Pleistocene while monsoon forests grew along the mountain front and streams. A tree savannah environment must have existed before the middle Pleistocene folding as indicated by the abundance of large grazing mammal fossils. The present vegetation patterns have been modified by agricultural practice, but tree savannah and grasslands appear to be the dominant vegetation with monsoon forests along the mountain front and streams.

The Cagayan Valley anticlinal belt formed in the middle to late Pleistocene (Figure 12). Accelerated uplift of the Cordillera Central formed a slope which made it possible for gravity sliding and asymmetrical-to-overturned folds to form. The resulting folds led to changes in the drainage system along the mountain front. The formation of Pangul Anticline restricted stream flow to the east as indicated by the series of rock cut terraces along the Tabuk plateau which documents the progressive diversion of the paleo-Chico River to the north. The terrace levels probably reflect uplift of the Cordillera and renewed downcutting. When the Chico River diverted to the north, the alluvial fan complex to the south became inactive and was partially dissected by streams. Erosion of the anticlines has provided sediment which has been accumulating in the adjacent synclines since folding took place. Numerous flat lying dacitic ignimbrites and lahars along the Tabuk plateau (Figure 12) indicate that dacitic volcanism continued after the folding episode.

Additional archaeological investigations need to be initiated to determine the date when hominids arrived in the Cagayan Valley. If tools or hominid remains can be found _in situ_ in the folded sediments of the anticlines, a middle Pleistocene age is suggested by the association with middle Pleistocene vertebrates and tektites. More precise dates may be obtained from whole rock radiometric dating techniques and/or paleomagnetic studies. If hominids inhabited the Cagayan Valley during the middle Pleistocene, gravel sources for tools were easily accessible along the meandering and distal braided streams. If the arrival of hominids post-dates the folding, gravels would also have been easily found after the folding due to erosional processes which have produced the lag gravels on many of the anticlines. The age documentation of hominids in the Cagayan Valley is now dependent on archaeological excavations to find tools or hominid fossils *in situ*.

SUMMARY

1. The Cagayan Valley basin is a north-south trending interarc basin which formed in a back arc setting between the active Cordillera Central and the inactive Sierra Madre volcanic arcs of Northern Luzon. Tectonic and volcanic events have dominated the sedimentary history of the basin which began to form in the late Oligocene following the initial uplift of the Cordillera Central volcanic arc. Eight thousand m of marine sediments were deposited in the Miocene as the basin subsided in response to development of the Cordillera Central. Uplift of the region in the Plio-Pleistocene resulted in the transition from marine to deltaic and then fluvial sedimentation.
2. Approximately 1,200 m of Plio-Pleistocene sediment have been deposited in the central part of the Cagayan Valley basin. These sediments have been divided into the Plio-Pleistocene Ilagan Formation and the Pleistocene Awidon Mesa Formation. The Ilagan Formation is divided into two members, a Lower Member formed by 150 to 310 m of interbedded sandstone and mudstone and an Upper Member formed by 500 m of pebble conglomerates, sandstones, siltstones, tuffs, and claystones. The Awidon Mesa Formation conformably overlies the Ilagan Formation and is composed of 400 m of pebble to boulder conglomerates, tuff-breccias, tuffs, sandstones, and claystones. Tuff-breccias and interbedded tuffs from 2 pyroclastic complexes, the Liwan Pyroclastic Complex and the Tabuk Pyroclastics which are designated as marker beds.
3. The sediments are composed primarily of volcaniclastic detritus. The conglomerates range from granule to boulder polymictic conglomerates

which contain primarily porphyritic andesite and basalt clasts in a litharenitic to arkosic matrix. Sandstones of the Ilagan Formation are primarily litharenites and feldspathic litharenites characterized by a pyroxene heavy mineral association. Plagioclase and quartz are more abundant in the Awidon Mesa Formation sandstones, which range from feldspathic litharenites to arkoses and contain a hornblende heavy mineral association. The mudrocks are primarily claystones composed of smectite. Smectite-kaolinite or kaolinite rich clay minerals assemblages occur in some mudrocks, more commonly the younger mudrocks of the Awidon Mesa Formation.

4. Most of the volcaniclastic sediments are diagenetically altered. Sandstones have been altered by the partial to complete dissolution of framework grains and the precipitation of authigenic clay, zeolite, and calcite cements. The Ilagan sandstones are more highly altered than the younger Awidon Mesa Formation sandstones. The more extensive diagenetic alteration of the Ilagan Formation reflects primarily the increase in diagenetic alteration with increased burial depth and time. Many of the mudrocks, especially whose which lack epiclastic detritus and contain euhedral phenocrysts of plagioclase, hornblende, and volcanic quartz, are interpreted as bentonites.

5. Five major lithofacies are recognized in the Plio-Pleistocene deposits. The facies and respective depositional environments are: 1) the interbedded fine grained sandstone and mudstone facies: delta front distal bar and distributary mouth bar; 2) the lenticular cross bedded medium grained sandstone and siltstone facies: delta plain distributary channel, levee, and flood basin; 3) the polymictic conglomerate, trough

cross bedded sandstone and claystone facies: low energy fluvial channel and floodplain; 4) the clast-supported polymictic conglomerate and sandstone facies: high energy channel bar and gravel sheet; and 5) the massive matrix-supported pebble to boulder conglomerate, tuff-breccia, and tuff facies: volcanic mudflows and debris flows (lahars), pyroclastic flow (ignimbrite) and fall (tuff).

6. The Plio-Pleistocene sediments were derived from volcanic and plutonic igneous rocks in the adjacent volcanic arcs. The composition of the sandstones and distribution of major lithofacies indicates that the augite-rich Plio-Pleistocene Ilagan Formation was derived from the arcs bordering the southern portion of the basin. The Pleistocene sediments which contain higher percentages of volcanic quartz, plagioclase, and a hornblende heavy mineral association were derived from the Cordillera Central to the west. The coarsening upward of the facies and pyroclastic deposits of the Awidon Mesa Formation also indicate that the Cordillera Central became the principal source in the Pleistocene.

7. The Plio-Pleistocene sediments record a regression of the Pliocene sea as the region was uplifted and the basin filled with detritus from the surrounding volcanic arcs. A lobate high constructive delta complex prograded to the north as it received sediment from a large meandering stream system which flowed down the axis of the valley. In the Pleistocene geanticlinal uplift of the Cordillera Central, volcanic arc resulted in the formation of an alluvial fan complex along the western side of the valley. Concomitant plinian and peléan volcanic eruptions in the Cordillera Central produced pyroclast flows which spread across the fan and into the valley.

8. In the middle Pleistocene, a terrestrial vertebrate fauna migrated to the Cagayan Valley by at least one land-bridge. At this time, most evidence suggests that a land-bridge existed between Luzon and Taiwan. Further studies of the newly recovered vertebrate fossils should provide additional evidence to more conclusively locate the land-bridge.
9. The vertebrate fauna inhabited the Cagayan Valley alluvial plain between the paleo-Cagayan River and the distal portion of the alluvial fan. A tree savannah ecosystem prevailed on the alluvial plain which was inhabited by the elephants, Elephas and Stegodon, rhinocerous, carabao (bovidae), deer, pigs, crocodiles, and turtles.
10. Tektites of possible extraterrestrial or impact origin were deposited by distal braided streams and/or small meandering streams in the plain between the alluvial fan and paleo-Cagayan River. The limited stratigraphic occurrence, K-Ar date of 0.92 ± 0.17 M yr. B.P. and association with middle Pleistocene fauna suggest that they are middle Pleistocene in age and probably correlative with the 710,000 year B.P. tektites of Southeast Asia.
11. Paleolithic stone tools found in surface association with the middle Pleistocene tektites and fossil fauna suggest that hominids also migrated to the Cagayan basin in the Pleistocene. Additional archaeological investigations are needed to document the age of the tools. If tools are found in situ, a middle Pleistocene age can be inferred from the association with tektites and the vertebrate fauna. More precise dates may be obtained from whole rock radiometric dating techniques and/or paleomagnetic studies. Thus far no skeletal evidence of hominids has been found in the sediments.

12. The Cagayan Valley anticlinal belt formed in the middle to late Pleistocene. Accelerated uplift of the Cordillera Central formed a slope which made it possible for gravity sliding and asymmetrical-to-overturned folds to form in the central Cagayan Valley.

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